THE UNIVERSITY OF ALBERTA

# MOISTURE MIGRATION IN FROZEN SOIL

by

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A THESIS

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To Mom, Dad, Maureen, Anne, Michael, Therese and Tim

## ABSTRACT

An experimental study was conducted to examine the moisture migration characteristics of frozen Devon silt. The objective of these experiments was to evaluate the influence of moisture migration and redistribution within the frozen zone on the process of frost heave.

The results from relatively long term open and closed system tests (the length of testing ranged from 5 to 22 days) on completely frozen silt samples revealed that moisture does migrate through frozen soil under the influence of a temperature gradient. It appears that the transfer of moisture occurs mostly within the unfrozen water film and that the development of ice lenses impedes the flow of water. Tritiated water was used as a tracer to help identify the location of the new water which had been drawn into the frozen soil.

It was also concluded from the test results that the process of regelation (movement of pore ice through a frozen soil) does occur as a result of an applied temperature gradient, but the effect on heave rate appears to be very small when compared with that of primary heave. Primary heave is defined here as moisture migration to the frost front as a result of a suction pressure within a frozen fringe of soil near the 0°C isotherm.

In most of the open system tests a definite drop in heave rate with time was noticed. This phenomenon was also observed in a field frost heave test conducted by others. Several interpretations are discussed in an attempt to explain this behavior.

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# CHAPTER I

## Introduction

### 1.1 Long Term Frost Heave

Frost heave is a phenomenon which is usually associated with the creation of segregational ice under roadways or other unheated structures. Although seasonal frost heave is important, far more destructive are the effects of a freezing soil-water system under the influence of a constant, long term step temperature (below 0°C). With the development of northern resources the construction of a "cold" natural gas pipeline seems inevitable. One of the more significant problems associated with the construction of such a pipeline is the effect of long term frost heave. This problem differs from seasonal frost heave in that the processes involved are imposed on a soil-water system continuosly over many years, although the processes themselves are expected to be the same. The time effects of frost heave have not been adequately examined to allow one to predict heave rates over any extended period of time.

It is recognized that water flows to the frost front and freezes into ice as a result of a temperature-induced suction force occurring somewhere within the frozen zone. The amount of segregational ice which forms depends upon the

soil type, step temperature (i.e.,frost penetration rate), overburden pressure, void ratio and the depth to the water table. Until recently, most investigations have been concerned only with the effects of these parameters on heave rate and heave pressures without examining the processes involved with frost heave. The ultimate purpose of most studies was to be able to predict frost heave rates in the field on the basis of laboratory tests. It was tacitly assumed by most authors that the data from short term laboratory tests could be extrapolated to predict long term field behaviour.

However, recent studies have shown that there are many processes involved with frost heave, at least some of which are expected to be time dependent. It has been suggested that moisture migration and redistribution within the frozen zone may contribute significantly to the overall heave rate. Simple, short term laboratory tests would not be effective in evaluating the long term influence of such processes on frost heave. It has become apparent that the understanding of these processes is essential for the development of a useful frost heave model.

## 1.2 Scope of the Thesis

It is considered important to examine in depth the processes associated with frost heave. As mentioned in Section 1.1, it has been suggested that one of these processes involves the migration of moisture through frozen soil. The object of this thesis is to investigate the migration characteristics of frozen Devon silt and to relate the results to the problem of long term frost heave.

A testing program was initiated to study various aspects of moisture migration through frozen soil. Closed system tests were conducted on frozen samples to examine moisture redistribution as a result of an applied temperature gradient. Open system tests were designed to simulate steady state field conditions. This represents the situation where the 0°C isotherm has stopped advancing (or slowed to a very low rate) but where water is continually being drawn into the soil where it freezes into ice. In the laboratory tests, uni-directional heat flow boundary conditions were imposed on a completely frozen silt sample. The sample was allowed access to water and subsequent changes in moisture content were measured.

From the test results, several conclusions are made regarding moisture migration characteristics of frozen soil and the significance of this process to frost heave.

# CHAPTER II

## Literature Review

### 2.1 Introduction

Most frost heave experiments prior to 1950 were designed to investigate basically whether or not a certain soil would heave. Gradually, more information was required for engineering projects and better designed tests were developed. Eventually, researchers realized that in order to adequately predict how a specific soil would react during freezing an understanding of the significant parameters affecting the process of frost heave was essential. Consequently, parametric studies were undertaken by Jumikis (1960) and Penner (1967), (1968), among others to examine the influence of pore size in the soil on frost heave rates. Penner (1960), (1972) and later Hill (1977) investigated the effects of temperature and temperature gradients on heave rates. Penner (1958) and Aitken (1963) provided quantitative evidence of Taber's (1929) observation that overburden pressure reduces heave rate.

During the same period of time other researchers were attempting to develop a model of frost heave based on known physical and thermodynamic relationships (e.g. Everett, 1961; Williams, 1966). The Thompson equation given in Equation 2.1 relates the size of a stable spherical crystal in its own melt and the absolute temperature.

$$Tr - To = \frac{2V \cdot \sigma iw \cdot To}{Riw \cdot L} \qquad (2.1)$$

where: Tr = freezing point of water (°C)
To = normal freezing point of water (0.0°C)
V = molar volume of water ($cm^3/g$)
$\sigma iw$ = surface tension of ice-water ($cal/cm^2$)
Riw = radius of ice-water interphase (cm)
L = latent heat of fusion (cal/g)

The Kelvin equation (Equation 2.2) relates the pressure difference between ice and water to the size of a crystal in its own melt.

$$Pi - Pw = \frac{2\,\sigma iw}{Riw} \qquad (2.2)$$

where: Pi = pressure of ice (kPa)
Pw = pressure of water (kPa)

By combining equations 2.1 and 2.2 a form of the Clausius - Clapeyron equation is obtained:

$$Pi - Pw = \frac{L \cdot (Tr - To)}{VTo} \qquad (2.3)$$

Everett (1961) proposed a theory of frost heave based on the above thermodynamic relationships. However, this theory and others (Takagi, 1977) are incomplete in that not all the physical processes of frost heave are taken into account. According to Hill and Morgenstern (1977) a complete theory of frost heave would lead to the prediction of:

a) variation of heave rate and heave pressure with time

b) water content (ice content) variation with time

c) temperature distribution with time

d) ultimate magnitude of heave

Until recently very few researchers have been concerned with the variation of frost heave processes with time. Nor have the physical processes themselves been adequately examined. It is important to understand exactly how ice forms in a soil before a model can be developed which would have the ability to predict the four points listed above.

It is the contention of this author that one important process of frost heave may consist of water migration within the frozen zone of the soil. It is accepted that water flows to the frost front as a result of a suction force (Beskow, 1935). Recently, Miller (1972) and a number of his students (Dirksen, 1966; Hoekstra, 1969a; Loch, 1975) have advanced the theory that the water is drawn past the zero degree isotherm and somewhere into the frozen zone. This phenomenon of water migration through the frozen soil may have major implications on frost heave prediction which will be discussed in the following sections of this chapter and in Chapter V.

The literature review is divided into three sections, each concerned with a specific aspect of moisture migration in frozen soil. Section 2.2 reviews pertinent studies involved with the properties of the unfrozen water film. Section 2.3 is concerned with the suction force developed

within this unfrozen water layer and its relation to frost heave. Finally, Section 2.4 treats the subject of actual moisture migration in frozen soil. Recent experimental evidence demonstrating this phenomenon is discussed. A summary of the more salient points obtained from the literature is presented in the final section of this chapter.

## 2.2 Unfrozen Water Film

When a fine-grained soil is frozen not all the water within the soil pores freezes at 0°C (Bouyocous, 1916; Lovell, 1957). Up to 50% of the moisture in a soil may exist as a liquid at temperatures of -2°C (Nersesova and Tsytovich, 1963). This unfrozen water forms what is termed a "film" around soil particles. The properties of this water film are very complex in nature and still not completely understood. But it is important to examine some of the better known properties of this layer of water before a discussion of its temperature-induced mobility is attempted.

The unfrozen water content is the amount of liquid water in a soil at a given temperature. It is usually expressed as a percentage of the dry weight of the soil, but in some cases it is expressed as a percentage of the total moisture content. The former definition will be used throughout this report. At any given temperature the moisture in a soil can be quantified as follows:

$$W = W_u + W_i \qquad (2.4)$$

where: $W$ = total water content (%)
$W_u$ = unfrozen water content (%)
$W_i$ = frozen water (ice) content (%)

The difference between the normal freezing point of water (0°C) and the temperature at which a certain amount of water is unfrozen is termed the freezing point depression, Tr, of the soil water. For a given temperature or freezing point depression, there exists a certain amount of unfrozen water. The relationship between freezing point depression and unfrozen water content has been examined by Williams (1964) and Anderson, et al (1973). Anderson presented an empirical relation in the form of Equation 2.5:

$$W_u = a \cdot T_r^{\,b} \qquad (2.5)$$

where: $T_r$ = freezing point depression (°C)
$W_u$ is defined in equation 2.4
a and b are constants dependent on soil type

Tice, et al (1976) examined the possibilities of obtaining values for a and b by relating them to the liquid limit of the soil. Preliminary results indicated excellent correlations with results obtained from adiabatic calorimeter determinations of unfrozen water content.

The freezing point depression, then, is directly related to the unfrozen water content. It is important to distinguish between freezing point depression and supercooling. The former is the temperature at which a certain amount of water and ice coexist in equilibrium.

Supercooling is a metastable condition where the same proportion of water and ice coexist but at a temperature lower than Tr (Wissa and Martin, 1968). This condition is rarely observed in the field but, due to a more controlled environment, it is commonly experienced in laboratory experiments (Anderson and Morgenstern, 1973).

Nersesova and Tsytovich (1963) presented evidence which showed that Wu is a function of temperature only and virtually independent of the total water content, In other words, if water is added to or taken from a soil at a constant temperature, the amount of water remaining as a liquid will be the same while the frozen water content will fluctuate accordingly. This implies that if water is being removed from an adiabatic system, some of the ice will melt in order to satisfy the equilibrium of the water film.

Nersesova and Tsytovich (1963) also presented data on the effects of exchangeable cations on the freezing point depression. In Montmorillonite, it was found that higher valence cations would reduce the amount of unfrozen water at a given temperature. However, little effect was observed when Kaolinite was tested using cations of different valence. This is to be expected since the thickness of the film is controlled by the amount of surface charge. Montmorillonite has a high surface charge (due to its high specific surface) and the thickness of its bound water is reduced by higher valence cations. Kaolinite has a smaller

surface charge and the film thickness would be less affected by the type of cation. Low, et al (1968b) reported that the effects of solutes is to increase the unfrozen water content as predicted by freezing point depression calculations.

Anderson and Hoekstra (1965a) used X-ray diffraction techniques to determine the thickness of interlamellar water at various sub-zero temperatures. The actual thickness of the interlamellar water in Wyoming Bentonite was found to be about 9 Angstroms (approx. 3 molecules in thickness) at a temperature of T = -6°C. Below this temperature the spacing decreased to 7 or 8 Angstroms and remained constant to -40°C. Ahlrichs and White (1962) obtained similar spacings down to liquid Nitrogen temperatures (-195°C) although Nersesova and Tystovich (1963) reported that in some Russian tests no liquid water was observed below -70°C. Anderson and Hoekstra (1965b) calculated a thickness of a water film in Montmorillonite to be on the order of 5 to 10 Angstroms at T = -10°C.

The structure of the unfrozen water has been examined by Anderson (1967). From X-ray diffraction tests he concluded that proposed Hydrogen bonding (by Hoekstra, 1965) of the interfacial water may not exist. However, the film does behave as a two-dimensional liquid with a structure different from that of free water. A preferred orientation of water molecules near a silicate surface was observed but the reasons for this are unclear. Tyutyunov (1963)

postulated the film water is a quasicrystalline body with geometric and structural components. He estimated that the attracting forces holding the water range from 1000 to 10,000 kPa.

Although the actual structure of the bound water is uncertain, it is generally agreed that the film possesses different properties than that of free water. The viscosity and density of the film are greater than in the case of free water (Anderson, 1967, Low and Lovell, 1959, Anderson and Low, 1958). Anderson (1967) concluded that the absorbed layer has some deformation properties, e.g., it has the ability to transmit a certain amount of shear. Anderson and Hoekstra (1965b) suggested that since solutes and other impurities tend to be excluded from ice crystals, the film may contain more solutes than normal thus increasing its structure. The interfacial water may also possess a latent heat of fusion less than that of normal water (Anderson, 1967).

The amount of unfrozen water at a certain temperature, T, was found to be dependent on whether the frozen soil was being warmed or cooled (Leonards and Andersland, 1960; Williams, 1963; Anderson and Hoekstra, 1965a). This hysteresis effect has not been fully explained but it does appear that the lower Wu associated with a cooling system may be attributed to the fact that the ice-water phases are not in equilibrium during cooling. It is expected that, in

time, the amount of unfrozen water in the cooling system would increase at a constant temperature up to or near the value of Wu for a warming system.

Anderson and Hoekstra (1965a) found evidence that ice forms as small crystals in the pore spaces and when the temperature is lowered, the crystal grows by extracting water from the film after the free water has been used up. It is of interest to discuss the actual location of the unfrozen water. Anderson, et al (1974) showed that in Montmorillonite, most of the unfrozen water is contained in the interlamellar regions (between clay particles) due to the fact that 80-90% of the surface area is exposed to these internal regions. Nearly all the surface area of Kaolinite, on the other hand, is exposed on external surfaces. Thus, more unfrozen water is retained per unit of total surface area in Kaolinite than in Montmorillonite, albeit, it is retained less tenaciously by the former clay mineral. They concluded that the unfrozen water interphase is thicker at a given temperature on external surfaces than on internal surfaces.

The authors, therefore, proposed a theory that various domains exist in a clay/water system. These domains contain water layers of various thicknesses. The water in the larger domains will freeze first drawing water from other areas of the system. This observation may be related to some results of Williams (1964). At a given temperature, Williams found

that the unfrozen water content for soils that had been previously frozen differed from that observed for the same soil during the first freezing. This implies that the structure of the soil system has some effect on the amount and distribution of unfrozen water.

The effect of pressure on unfrozen water at a constant temperature is to increase the unfrozen water content (Anderson, 1967; Low, et al, 1968b). This will be discussed in greater detail in the next section on suction forces (Section 2.3).

From the above discussion it can be concluded that the unfrozen water is a somewhat structured film with properties different from those of free water. It is generally agreed that the structure and thus viscosity becomes more pronounced as the distance to the particle surface decreases.

The thickness of the film is dependent upon many factors, the most important of which are listed below:

a.) soil type (specific surface area and surface charge density)

b.) temperature

c.) configuration of individual soil particles

d.) quantity and type of exchangeable cations

e.) overburden pressure

The first three factors are considered to be of major

importance in the formation of unfrozen water while the last two are usually much less important (Anderson, et al, 1974).

## 2.3 Suction Forces in Frozen Soil

The fact that water is drawn to the freezing front due to a suction within the soil is well documented (e.g. Taber, 1929). Though not as well documented, it is becoming generally accepted that water migrates through frozen soil as a result of these same forces (Miller, 1972; Williams, 1977; Penner, 1978). This section concerns itself with the nature of the suction and the effects of temperature and pressure on its magnitude.

Suction can be induced in a soil by various methods. Regardless of its nature, each method produces a suction in the soil by causing a difference in potential from one part of the soil to another. Jumikis (1966) discusses four major causes of potential difference:

a) thermal

b) electrical

c) capillary

d) osmotic

The net effect of any applied potential difference in a soil is to produce a potential gradient within the soil water. On a microscopic scale, it is suggested that the gradient leads to a lowering of the Gibbs free energy in the

colder section of the film layer near the soil surface (Harlan, 1974; Williams, 1977).

Williams (1963) presented experimental data along with a theoretical relationship between temperature of the soil and suction. This is given in Figure 2.1. Williams obtained the experimental values by relating suction in an unfrozen soil at a certain degree of saturation (less than 100%) to a frozen soil with the same quantity of film water as in the unfrozen soil. As mentioned before, there exists a direct relationship between unfrozen water content and temperature; thus, Williams could indirectly relate temperature to suction. The difference in surface energy for an ice-water and air-water interphase was taken into account in the preparation of Figure 2.1. The ratio between the respective interphase energies, Oiw and Oaw, was taken to be about 0.4, respectively (Williams, 1966; Hesstvedt, 1964).

Harlan (1974) presented a relationship between Gibbs free energy (or suction) and temperature based on fundamental thermodynamics. His relationship, given in equation 2.6 is formulated from the same physics as Equation 2.3.

$$G = - \frac{RT}{M} (\ln P/Po) \qquad (2.6)$$

where: G = soil water potential
T = temperature of soil
R = gas constant
M = molecular weight of water
P = vapor pressure of soil water
Po = vapor pressure of bulk water at T and P

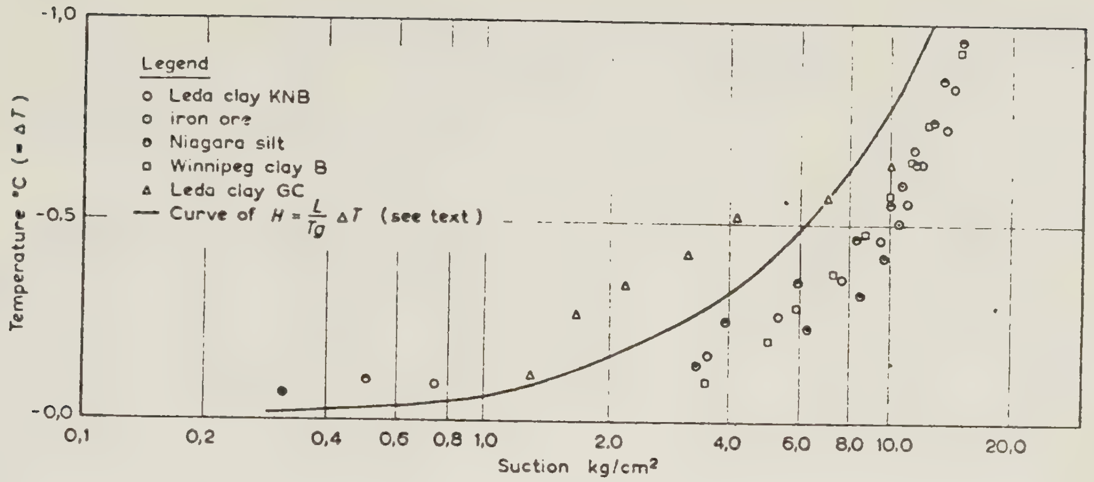



FIGURE 2.1 THEORETICAL AND EXPERIMENTAL RELATIONSHIP BETWEEN SUCTION AND TEMPERATURE.
(AFTER WILLIAMS, 1963)

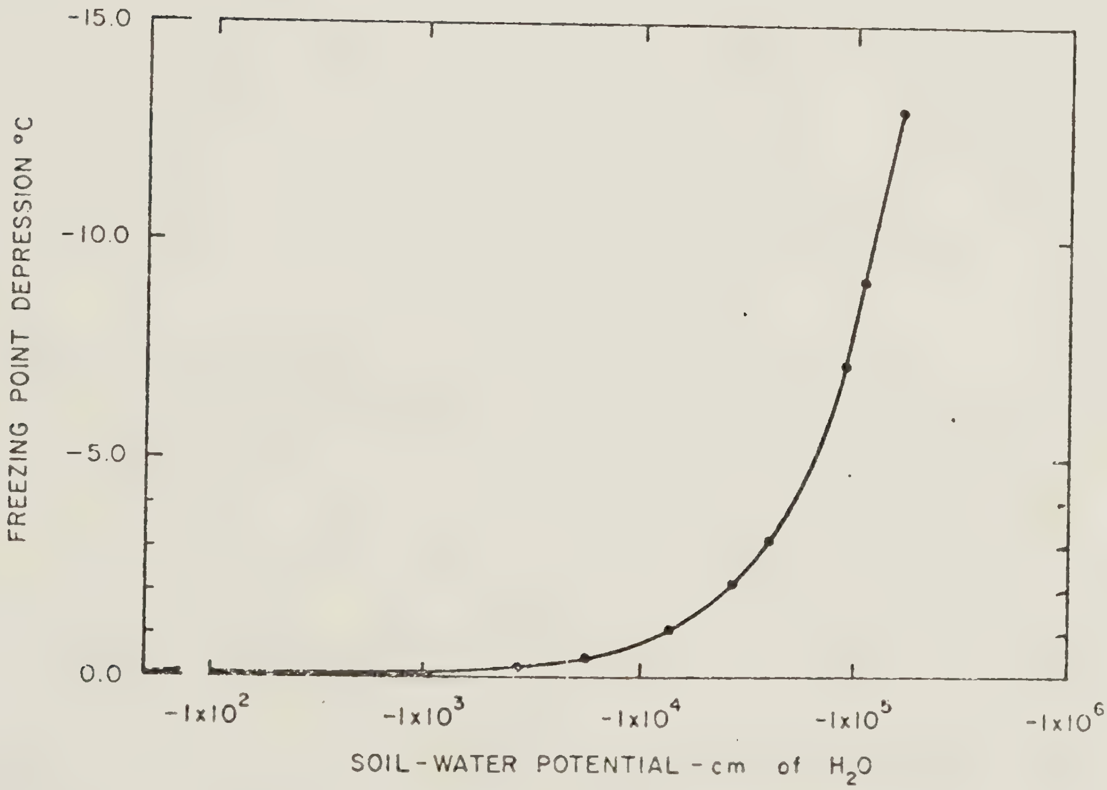



FIGURE 2.2 THEORETICAL RELATIONSHIP BETWEEN FREEZING POINT DEPRESSION AND SOIL-WATER POTENTIAL. ( AFTER HARLAN, 1974 )

Equation 2.6 relates theoretically the freezing point depression to the soil-water potential. This is shown graphically in Figure 2.2.

Since the suction in a frozen soil is due to the lowering of the water potential (or freezing point depression), from Equation 2.3 it follows that the application of pressure to the system would influence the negative pressure in the soil water. Wissa and Martin (1968) presented an excellent discussion of the effects of pressure on freezing point depression. They concluded from examination of pertinent research (Beskow, 1935; Miller, et al, 1960; Williams, 1963) that, under application of a load, the pressure on the ice remains constant while the pressure on the water changes according to a modified form of Equation 2.3:

$$\frac{dT}{dPw} = \frac{V \cdot To}{L} \qquad (2.7)$$

where: dPw = change in pressure on water (kPa)
dT = freezing point depression (°C)
(the remaining symbols are as previously defined)

According to Equation 2.7, decreasing the water pressure by 100 kPa lowers the freezing point by 0.0803°C (and vice versa). However, when dealing with a real soil system the authors report that cavitation of water will occur at a tension of about 75 kPa.

This corresponds to a freezing point depression of 0.06°C. This temperature at which cavitation supposedly

occurs appears to be extremely warm in light of experimental studies presented in the following section which demonstrate moisture movement in a soil at temperatures near -2.0°C (e.g. Kudryavec, et al 1973). It is suggested that the present state of knowledge regarding cavitation in a frozen soil-water-ice system is not at a point where an adequate understanding of its influence on moisture migration in frozen soil exists.

Many experiments have been conducted to study the origin and magnitude of heave pressures. Taber (1929), Penner (1959), (1963), (1967a), Miller, et al (1960), Hoekstra, et al (1965a) and Hoekstra (1969b) all performed tests to evaluate heave pressures. The tests were similar in that the movement of the soil due to frost heave forces was restricted by pressure measuring devices; i.e. very little or no movement was allowed. Pressures up to 1,400 kPa and higher have been measured in fine-grained soils in the laboratory (Hoekstra, 1969b). Penner (1970) conducted a field frost heave test on Leda clay. Using a rigid, non-yielding plate equipped with a load cell to measure heave pressures, Penner obtained values up to 1,800 kPa. The pressure on the plate, however, was not due only to heave of the soil directly below the plate but, in addition, to heave pressures from soil in the general area. The method of pressure transfer from heaving of the outlying soil was considered to be arching due to soil stiffness.

It is felt that these types of tests are not indicative of actual frost heave pressures developed under normal circumstances where a soil is allowed to heave with only modest overburden pressures applied. A probable reason for such high pressures under a zero-movement condition is that the growth of an ice lense was impeded. This situation would not be expected to occur under normal loads. It is expected that a significant proportion of the heave pressure measured in the above tests could be attributed to the volume change of water to ice.

Arvidson and Morgenstern (1974) and later McRoberts and Nixon (1975) proposed the concept of "shut-off pressure". This pressure is defined as "the stress at the frost line which will cause neither flow of water into or away from the freezing front". Short-term frost heave tests (several hours) by Arvidson (1973) and Hill (1977) indicated that shut-off pressures exist for a given frost suseptible soil. However, longer frost tests (up to several days) by Hill (1977) and by Penner and Ueda (1977) showed that under high overburden pressures of 400 kPa, water was initially expelled from the soil but after about one day the process was reversed and water intake, and subsequently frost heave, commenced under the same load. No attempt in either testing program was made to study the effects of even higher pressures, probably because it would bring the test conditions out of the realm of normal engineering design.

These latter tests, then, introduced some doubt as to whether a shut-off pressure does, in fact, exist for a soil water system.

The results of the above long term heave tests also showed that the heave rate decreased slightly with time. No adequate explanation was offered in either case.

In experiments involving the effects of pressure and temperature on the growth of ice in an ice/quartz filter/water system, Biermans, et al (1976) and Biermans, et al (1978) found what appears to be a shut off pressure. Their test set-up consisted of a large ice crystal (diameter = 7 mm) over a thin quartz filter (pore size = 50 nm) which separated the ice from super-cooled water. Various temperatures and pressures could be applied to either the ice or the water. Results showed that, at a certain temperature, ice growth would occur with an increase in water pressure (or, likewise, decrease in ice pressure). When the water pressure was decreased to a certain point, the growth of the ice crystal stopped and further decrease in water pressure caused the ice to melt. Moreover, the behavior was in accordance with that predicted by the Clausius-Clapeyron equation (Equation 2.3). Vignes and Dijkema (1974) explained the results by suggesting that the suction force is developed in the thin, unfrozen, absorbed water layer between the ice and the quartz filter.

The fact that the relation between temperature and pressure in these experiments follows Equation 2.3 to a much higher degree than in a frozen soil-water system is an important point. It is suggested that a reason for this could be the difference in the properties of the absorbed film layer between the two systems. The water film within the ice-quartz filter interphase would be held less tenaciously due to the relatively low activity of quartz. The bonding forces in quartz are derived from the ends of broken tetrahedral crystals resulting in a very low potential to attract water. Thus, the water layer would be expected to have little structure and should not possess a very large potential difference from that of free water. Therefore, the water film would be more easily affected by temperature and pressure. In a soil-water system, however, the unfrozen film is strongly attached to the particle surface. The actual forces involved within the film are uncertain, but it is thought that the cumulative effects on the properties of the film would be far greater than in the simple filter-ice-water system.

In addition, the unfrozen film in soil is far more extensive than is the single layer between the ice and quartz filter. A somewhat continuous network of interconnecting films exist in frozen soil. The extent of unfrozen water is present throughout the frozen zone. Therefore, the effects of temperature and pressure

variations are distributed over the entire frozen zone. Actual distributions are unknown at this time. The significance of this point is that the applicability of thermodynamics to model frost heave depends upon how well the processes within the frozen zone are understood. In a soil system complicated, interrelated phenomena make thermodynamic modelling difficult. The processes are also probably time dependent introducing an additional parameter to be included in the analysis.

The tests of Hill (1977) and Penner and Ueda (1977) were designed to investigate the influence of pressure and temperature on frost heave rates. More recently, Penner and Ueda (1978) proposed an empirical relation to connect the three parameters with one equation:

$$dH/dt = a \cdot e^{(P/T)} \qquad (2.8)$$

where: $dH/dt$ = total heave rate (mm/min)
P = overburden pressure ($kg/cm^2$)
T = cold-side temperature (°C)
a and b are constants (function of soil type)

Results from tests on various soil types are shown on a log-log plot in Figure 2.3.

In their paper, the authors concluded that heave rates are affected by the cold side temperature only since no correlation between the rate of frost penetration and heave rate was found. It seems apparent to this author that the two are related; the rate of frost penetration is determined

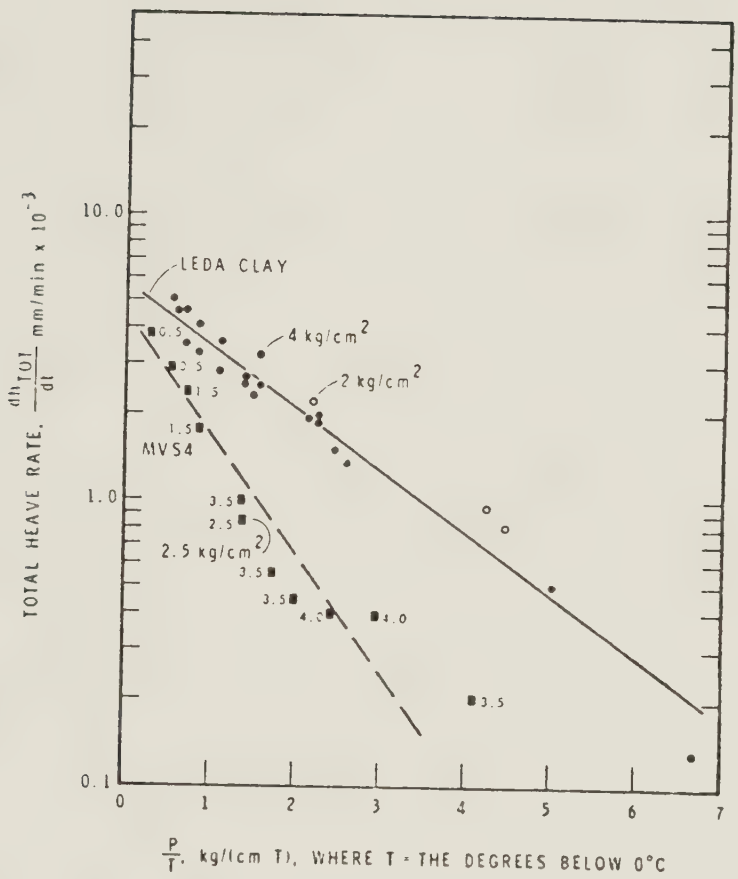



FIGURE 2.3 THE RELATIONSHIP OF HEAVE RATE TO OVERBURDEN PRESSURE AND STEP TEMPERATURE.

(AFTER PENNER AND UEDA, 1978)

by the temperature of the cold end of the sample. It is suggested that the lack of correlation results from the non-steady state condition during freezing. The rate of heave is controlled by the temperature gradient at a microscopic level which is affected by particle configuration, ice configuration, flow of water, etc., and is nearly impossible to identify accurately. During steady state conditions, i.e., when the frost front has stopped advancing, it would be expected that the temperature gradient in the frozen zone would be a controlling factor on the rate of heave. This is discussed further in Chapter V.

Penner and Walton (1978) proposed a method of analysis to include the effect of overburden pressure on the rate of ice accumulation within the frozen soil. Based on semi-empirical relations only (by differentiating Equation 2.8 with respect to temperature) they obtained correlations which indicated that higher pressures will cause the rate of ice accumulation near the zero degree isotherm to decrease and to cause faster accumulation deeper within the frozen zone. Examples of this are given in Figure 2.4 and 2.5. Also undertaken by Penner and Walton, was an explanation for the observed decrease in heave rate with time. The authors introduced "ice segregation ratio" which is the ratio between the total amount of heave and total frost penetration. Assuming that heaving is uniformly distributed over the frozen zone they presented the relation:

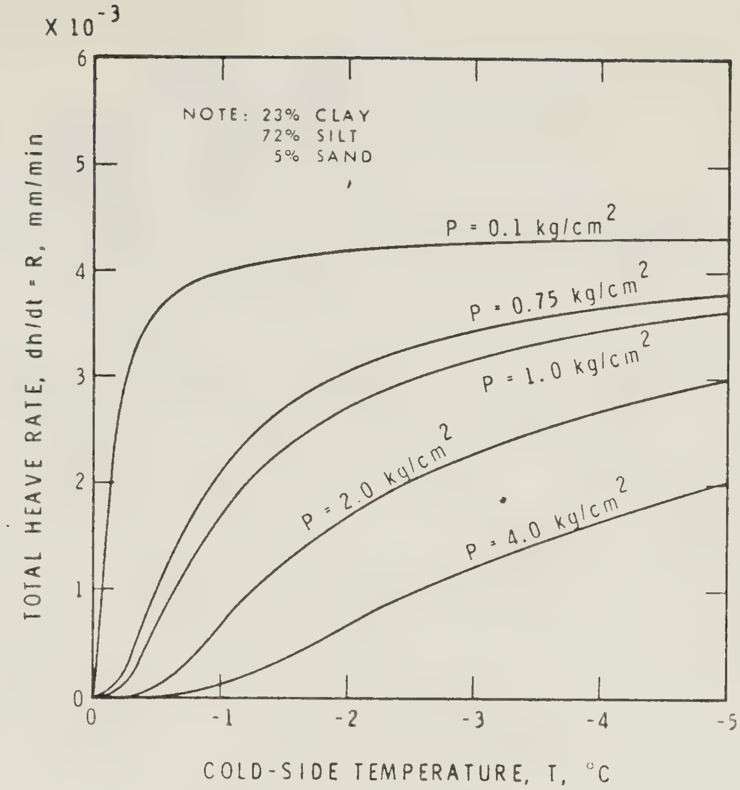



FIGURE 2.4 ICE ACCUMULATION RATE VS. COLD SIDE TEMPERATURE FOR SILT.
(AFTER PENNER AND WALTON, 1978)

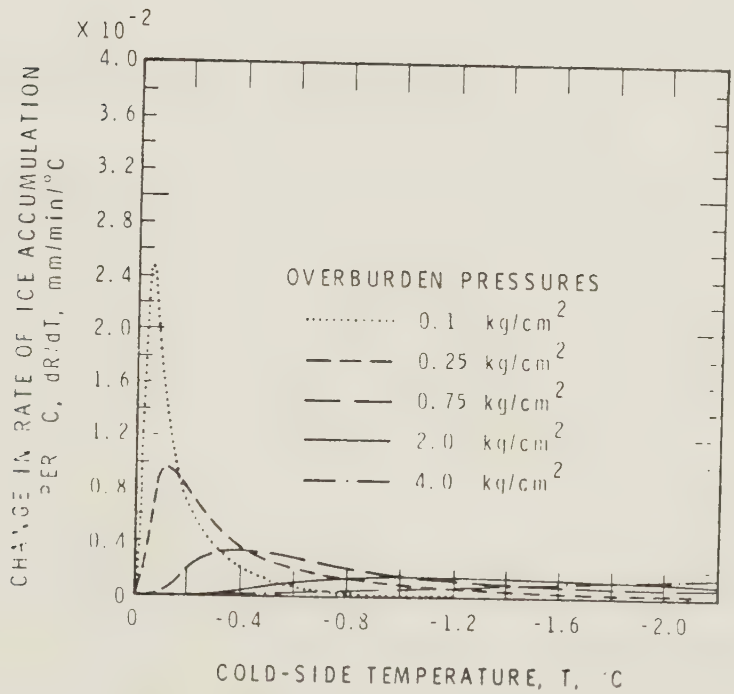



FIGURE 2.5 CHANGE IN RATE OF ICE ACCUMULATION PER DEGREE CELSIUS VS. COLD SIDE TEMPERATURE FOR SILT.
(AFTER PENNER AND WALTON, 1978)

$$dH/dt = Ro(1 - h/x) \qquad (2.9)$$

where: $dH/dt$ = total heave rate (mm/min)
$h$ = heave (mm)
$x$ = frost penetration (mm)
$Ro$ = intial heave rate (mm/min)
$h/x$ = ice segregation ratio

The correlation with actual lab tests on Mackenzie Valley silty clay was acceptable for a relatively short period of time. The calculated heave rates (based on Equation 2.9) at the end of the experiments, the longest of which lasted 12.5 days, was consistently higher than the actual heave rates. It was postulated that the reason for this discrepancy was that a solid ice lense prevented water from migrating into the frozen zone. Thus, the simplifying assumption associated with Equation 2.9 is questionable.

## 2.4 Moisture Migration in Frozen Soil

From the discussion on unfrozen water content and suction forces it seems reasonable to postulate that the unfrozen water is mobile and can migrate through the frozen zone under the action of a temperature-induced (or other) potential gradient.

In 1964, Hoekstra and Chamberlain conducted electro-osmosis tests on frozen Wyoming Bentonite and New Hampshire silt in a closed system. The results are presented in Table 2.1 below. An electric gradient of 1 Volt/cm was applied for a duration of 24 hours.

TABLE 2.1

Results from Electro-Osmosis Tests on Frozen Soils

| Soil Type | Temperature (°C) | Initial W/C (%) | Final W/C (%) 1 cm from Anode | Final W/C (%) 1 cm from Cathode |
|---|---|---|---|---|
| Bentonite | -2.0 | 341 | 275 | 456 |
| | -1.5 | 265 | 134 | 310 |
| Silt | -1.5 | 30 | 27 | 32 |
| | -1.0 | 28 | 21 | 40 |

The test results revealed that a considerable quantity of water was transported in frozen soil towards the cathode. The authors noted that in the final state all the ice from the anode region was removed and large bodies of ice were formed in the vicinity of the cathode. Shrinkage cracks were observed near the anode.

The mobility of the water film and the effect of temperature on the degree of mobility was adequately demonstrated in the above tests. The fact that flow was induced by the application of an electric gradient should not be unexpected. Vershinin, et al (1949) showed that the conductance of frozen ground was higher than anticipated for either dry soil or ice. He concluded that a continuous water film around the soil particles must exist. Hoekstra and Chamberlain attributed the high volume of the electrically-induced flow to an excess of cations in the film water due to the fact that solutes and other impurities are excluded

from the ice phase.

Hoekstra (1965) performed electrical conductance tests on frozen Wyoming Bentonite. From the results, he concluded that:

1) conductance per gram of frozen soil is independent of total water content
2) at low spacings, mobility of the film is almost independent of ionic species
3) mobility may be controlled by the degree of Hydrogen bonding
4) films are continuous since the conductance from tests coincides to an adequate degree with that predicted by theory.

The actual measurement of hydraulic conductivity, or permeability, in frozen soil was undertaken by Williams and Burt (1974). They used a special device in which a frozen soil sample was separated from lactose solution baths on either side by means of a 'wettable' porous membrane. A pressure was applied to a lactose bath on one side at various temperatures and the flow into the other bath was measured. Permeabilities based on Darcy's Law could then be calculated. Typical results are given in Figure 2.6. A major drawback to this method was that the membrane pores were too large to prevent lactose molecules from passing through. Thus, one side of the soil became saturated with lactose causing melting of ice in this region. Some consolidation

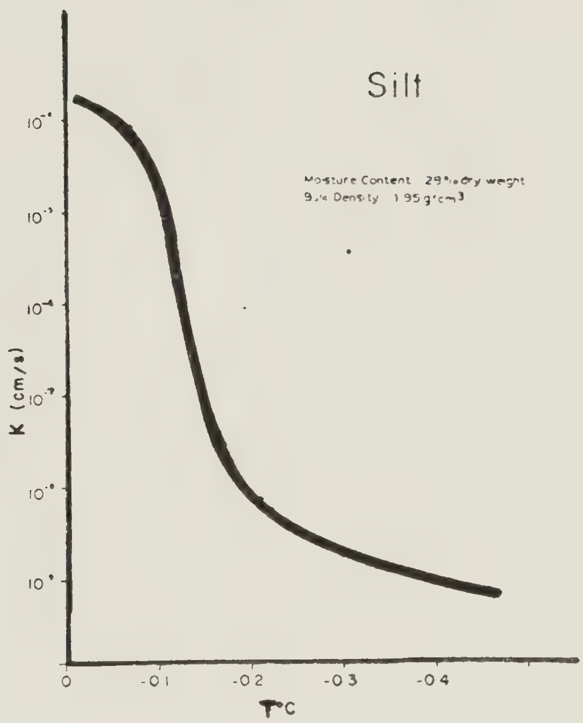



FIGURE 2.6 HYDRAULIC CONDUCTIVITY OF FROZEN SILT AS A FUNCTION OF TEMPERATURE. (AFTER WILLIAMS AND BURT, 1974 )

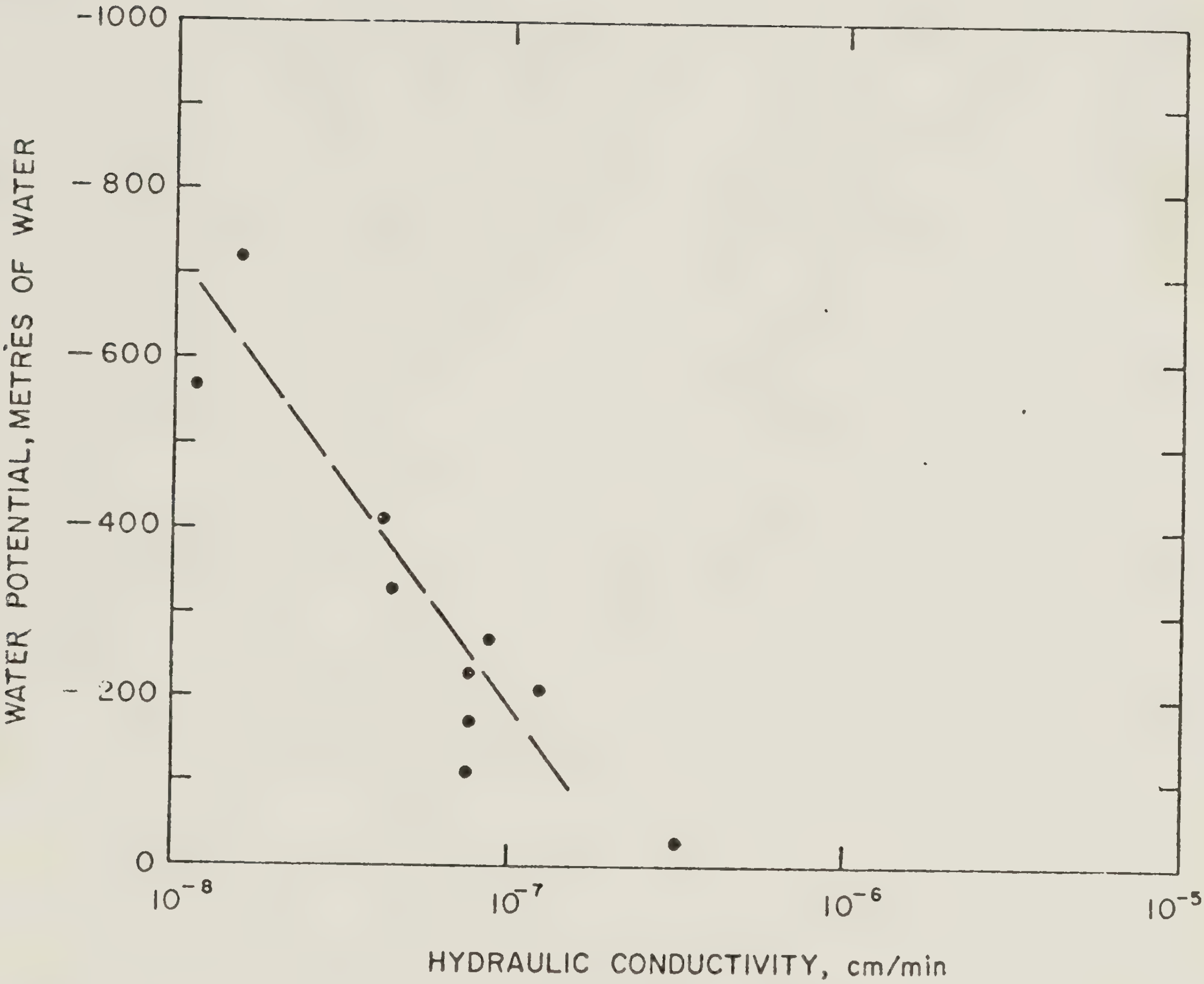



FIGURE 2.7 CALCULATED HYDRAULIC CONDUCTIVITY AS A FUNCTION OF WATER POTENTIAL.
— DATA ORIGINALLY FROM HOEKSTRA, (1966)
( AFTER HARLAN, 1974 )

could have occurred as a consequence of this partial thawing and application of pressure which would result in the generation of some excess pore pressure. The significance of this effect was not examined by Williams and Burt.

Harlan (1974) discussed the flow of moisture in natural frozen ground. He mentioned other factors besides temperature that would influence migration rates in the natural environment such as the distribution of ice lenses, quantity and type of solutes, configuration of cracks, etc. He also presented the frozen permeability, Kf, as a function of water potential. This is shown in Figure 2.7. The frozen hydraulic conductivities presented in Figure 2.7 were calculated from the observed rates of flow and thermal gradients in Frozen Fairbanks silt reported by Hoekstra (1966). From these values of permeability Harlan estimated rates of moisture migration in this material to range from 0.05 to 11 cm/year.

Several tests have been conducted to evaluate the migration rates of liquid and solid inclusions in single crystalline ice. The purpose of discussing these tests is to show the mobility of particles through ice, thereby providing a basis for postulating that ice and water, likewise, are mobile in a frozen soil. This will be made clearer in the following discussion.

Hoekstra, et al (1965b) tried to relate actual

migration rates of a liquid brine pocket with those predicted on the basis of diffusion theory. They obtained migration rates of 50 to 150 microns/hr at temperatures in the range of 5 to 15°C. These values coincided with that predicted by diffusion only when the assumption was made that the thickness of the unfrozen water film is on the order of 100 Angstroms. This estimate far exceeds the thickness of the film measured by Anderson and Hoekstra (1965a).

Hoekstra and Miller (1967) looked at the effects of temperature and temperature gradient on the movement of glass beads through ice. They obtained migration rates of 0 to 6 microns/hr at a temperature gradient of 1.0°C/cm. For a constant gradient, the higher velocities were associated with colder temperatures. They explained the movement of the glass beads on the basis of diffusion of the unfrozen water on the glass surface. Large values for film thickness were calculated using the diffusion model only.

Romkens and Miller (1973) conducted experiments involving the movement of glass beads and some minerals through ice. Rates of migration were an order of magnitude higher than those found by Hoekstra and Miller (1967). Velocities were not systematically related to particle size or temperature gradient and, in fact, some of the particles did not move at all. Also, there was no notable difference between tests with distilled water and those with a NaCl

solution. Since the diffusion model previously used called for film thicknesses far in excess of those observed, two alternate models were proposed. Both models use osmotic transport of water in the film as the driving force for migration.

It is seen from the above tests that the development of a theory to predict behavior at a microscopic level is difficult, at best. These tests are important, though, because they do provide more insight into the behavior the unfrozen film under the influence of a temperature gradient.

The actual migration of ice and/or water in frozen soil behind the freezing front was observed by Dirksen and Miller (1966). In closed system tests on unsaturated, New Hampshire silt they found that the water content of the unfrozen soil adjacent to the 0°C isotherm decreased with time. Moreover, the water content increased over an extended region behind the frost front with time, obviously drawing water from the unfrozen soil. In several samples no ice lense formation (or heave) had occurred. A higher degree of moisture redistribution was generally noticed in these samples. This was explained by the fact that ice lenses may tend to inhibit the movement of water within the unfrozen film by causing breaks in the continuity of the films. This explanation was also postulated by Hoekstra (1969b).

Hoekstra (1966) found similar results as those above

for unsaturated Fairbanks silt. In open system tests, Hoekstra measured the accumulation and redistribution of water content by means of gamma-ray attenuation. He also found that ice lenses formed a considerable distance behind the frost front. Ice lense formation continued with time within a certain zone long after the 0°C isotherm had passed through the zone. Both Dirksen and Miller (1966) and Hoekstra (1966) dismissed the possibility of vapor transfer as a plausible explanation for the amount of moisture transfer observed. They concluded that, based on calculations by Philip and deVries (1957), the amount of moisture movement possible as a result of vapor transfer would be on the order of 1000 times less than that observed. Hoekstra concluded that moisture movement through frozen soil may be very significant in the process of frost heave.

Several Russian authors have studied moisture migration in frozen soil. Ershov, et al (1976) investigated the migration and redistribution of moisture in the frozen zone of thawing soils. The experimental setup consisted of 4 X 4 X 15 cm clay sample covered with foam insulation and placed in an apparatus wherein various temperatures could be applied. Below freezing temperatures (-1°C to -15°C) were applied to one end of the sample while above freezing temperatures (1°C to 10°C) were applied to the other. The sample was allowed access to water from the warm end. Temperatures were monitored along the sample and changes in

the ice structure were recorded by time-lapse photography. The tests lasted up to 6 days.

Results from tests on Kaolinite clay showed that ice bands of various thicknesses developed within the frozen zone of the thawing sample. The ice lenses became thicker toward the thaw front (up to 2 cm near the 0°C isotherm). It was also observed that as the thaw front progressed, the ice that was melted appeared to migrate into the frozen zone and served as a source of supply for new ice seams forming in the frozen zone. Subsequently, a decrease in water content in the thawed zone and related consolidation was noted. The results of tests on Montmorillonite were similar except that the magnitude of lensing was less. Additional tests in soils with an initial massive ice structure showed similar phenomenon with the initial ice lenses increasing in thickness near the thaw front.

These tests provide important evidence that the migration of moisture into frozen soil does not depend on the magnitude or direction of heat flux. It appears that the suction force is only a function of the temperature gradient in the frozen soil. This perception could be extended to field conditions where massive ice lenses have been observed near the surface of a degrading permafrost environment. Chatwin (1978) noted such ice structures in the Mackenzie Valley area near Fort Simpson. Tritium dating placed the date of this near surface ice in the order of 100 years.

However, carbon dating of the peat in which the ice is contained resulted in typical values of 4000 years. Therefore, the process of moisture migration in frozen ground from the surface appears to be a possible mechanism for the growth of near surface ice lenses.

Kudryavec, et al (1973) performed very interesting closed system tests for several clay soil types. Three test configurations were used.

The first set of tests consisted of placing a pre-made ice lense on the bottom (warm end) of a completely frozen sample. The preparation of the sample consisted of compacting a slurry in a 4 X 4 X 2 cm mold and quick freezing. A temperature gradient of 0.6°C was then applied with no moisture transfer to the outside allowed. The migration of moisture within the sample was measured by taking water contents of thin slices of the sample after completion of the tests. The duration of the testing ranged from 4 to 20 days. The results are shown in Figure 2.8 (a). It is seen that part of the premade ice lense had migrated to the center of the sample where a new ice lense was formed. It was noted by the authors that the initial ice lense had developed a weaker or more porous structure during testing. The results provide good evidence that water can migrate through the unfrozen film with the application of a temperature gradient.

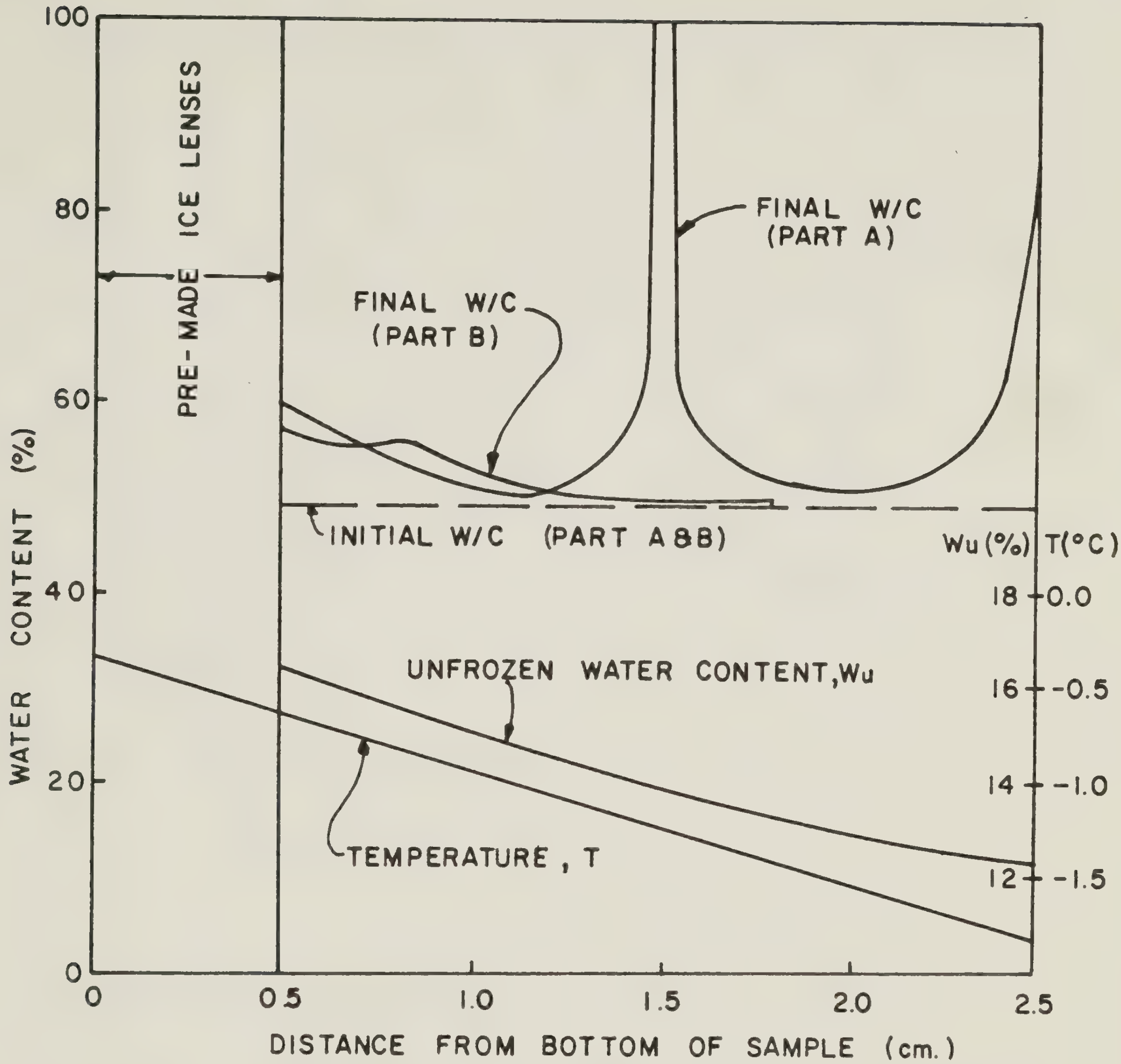



FIGURE 2.8 : (A) RESULTS FROM CLOSED SYSTEM TESTS S ON FROZEN MONTMORILLONITE WITH A PRE-MADE ICE LENSE. TEMPERATURE GRADIENT = 0.6 °C/cm. TIME = 14 DAYS

(B) WATER CONTENT PROFILE FOR TESTS WITH NO APPLIED TEMPERATURE GRADIENT TIME = 8 DAYS

(FROM KUDRYAVEC, et al, 1973)

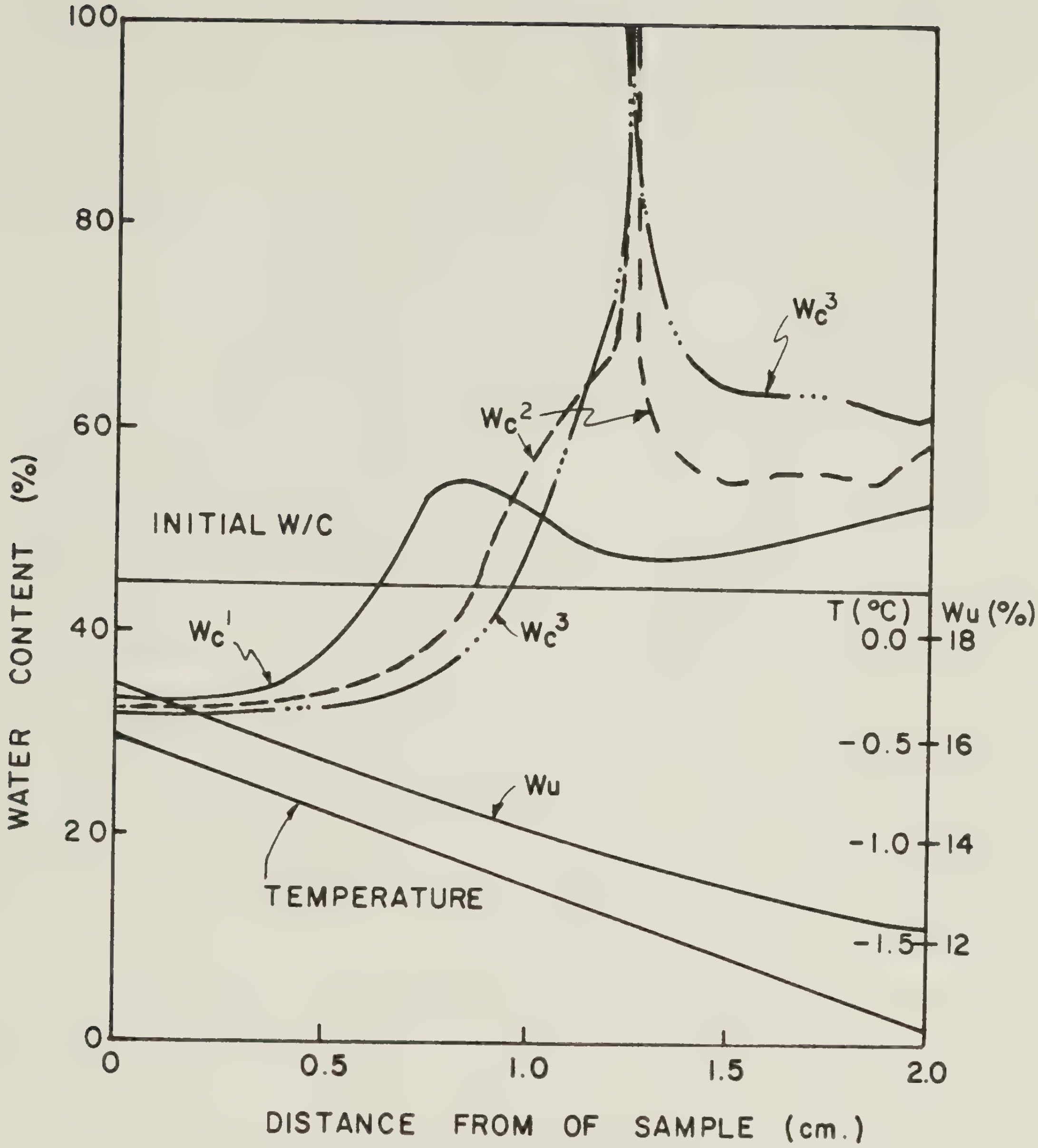



FIGURE 2.9: RESULTS FROM CLOSED SYSTEM TESTS ON FROZEN MONTMORILLONITE WITH NO PRE-MADE ICE LENSE

———— $W_c^1$ = FINAL WATER CONTENT AFTER 4 DAYS
– – – – $W_c^2$ = FINAL WATER CONTENT AFTER 8 DAYS
—···—··· $W_c^3$ = FINAL WATER CONTENT AFTER 12 DAYS
$W_u$ = UNFROZEN WATER CONTENT

( FROM KUDRYAVEC, et al, 1973 )

A second series of tests were conducted in the same manner as the first except no temperature gradient was applied. The results from a typical test are shown in Figure 2.8 (b). A slight increase in water content was observed in the frozen soil adjacent to the pre-made ice lense. The authors explained this migration by suggesting that a greater energy exists within the unfrozen water than in the ice lense. It is this potential difference, they conclude, which causes regelation of the ice lense and subsequent moisture migration into frozen soil.

The set-up for the third set of experiments was similar to the first except that the pre-made ice lense was deleted. Under a temperature gradient of 0.6°C/cm a redistribution of moisture was observed. The results shown in Figure 2.9 reveal that, as in the first set of tests, an ice lense developed near the middle of the sample. Desaturation in the warmer zone of the frozen soil caused localized stresses to develop which led to irregular cracking of the soil. In addition, the cracking induced discontinuities in the flow lines for moisture migration; thus, the process of moisture movement gradually slowed down and eventually stopped.

Similar results were obtained using other soil types - only the magnitude of moisture migration differed. Also, it was observed that, in tests with intially unsaturated conditions, only after the moisture derived from the pre-made ice lense had redistributed itself evenly through the

frozen soil did a zone of high water content develop.

In 1972 Miller proposed the concept of "Secondary Frost Heave" to help explain the effects of ice movement through the frozen soil. In the process of secondary heave ice is transported towards the colder side of a sample by means of a regelation process in which a constant melting at one end and subsequent refreezing at the other end of pore ice occurs. Miller based this theory on some of the literature mentioned before and on the results one of his earlier experiments. In 1970 Miller conducted a test to examine mass transport of ice due to a temperature gradient in a device he called an "ice sandwich". This device which was actually a type of osmometer consisted of a container in which an ice layer was "sandwiched" between salt water and a body of distilled water, separated by a porous filter on either side. Pressure could be applied to either the soil or the water. The system was similar in nature to that used by Williams and Burt (1974). The results of this test indicated that movement of ice occurs as described above. A value for the apparent permeability of ice, kw, based on a form of Darcy's Law ranged from $4 \times 10^{-9}$cm/sec at -0.07°C to about $2 \times 10^{-9}$cm/sec at -0.15°C.

Miller(1972) and Miller, et al (1975) described the driving force for the regelation of pore ice as the pressure distribution in the ice due to the presence of a temperature-induced potential gradient. The pressure

gradient in the unfrozen film is directly related to that in the ice since the ice and film are in equilibrium with one another. Thus, Miller, et al (1975) concluded that the "movement of pore ice can be a component of total transport whenever soil particles are separated from ice by an unfrozen film." Miller (1972) suggested the additional heave pressures developed from ice transport could be an explanation for the fact that Penner (1967) obtained consistently lower pressures than predicted by theory involving only primary heave. Penner based his calculations on the assumption that the driving force for water intake developed only at the freeze-thaw interphase. Miller termed this primary heave. It is suggested by this author that primary and secondary heave are related phenomenon. This point will be discussed further in Chapters V and VI.

One aspect of Miller's theory of secondary heave is the effect of pore ice configuration on the flow of water through the unfrozen film. According the Thomson equation (Equation 2.1) at temperatures near 0°C the smallest radius of the soil pore in which an ice crystal will form will be very large. The ice crystal in these pores would be expected to be spherical. Due to the nature of pores in a soil matrix the triangualar shaped areas adjacent to the ice crystal and the soil particles would be filled with water. At colder temperatures a tendency for the ice crystals to extend into these crevices would develop. The configuration of the

crystals would no longer be spherical but would conform somewhat to the shape of the pore. This situation would result in a net decrease in apparent permeability since the water would have to travel further to advance through the frozen soil.

Miller (1978) presented a refinement of his earlier theories on secondary heave and included a tenative solution which coupled mass transport in the frozen zone with mass transport to the frost front. In his derivation, Miller discussed several methods of moisture migration. Besides regelation of the pore ice other forms of tranport include moisture migration through the unfrozen film and through ice free pores. Miller, Loch and Bresler (1975) discussed the interaction of the various modes of transport and termed the summation of the effects "series-parallel transport". However, the notation was too difficult to be of any practical use. Miller (1978) circumvented this problem by simplifying the coupling of mass transport in frozen soil in the following relation:

$$Q = -kf \cdot (\partial T/\partial z) - \rho i (L \cdot Vi) \qquad (2.10)$$

where:
- Q = total heat flux
- kf = thermoconductivity of frozen soil
- $(\partial T/\partial z)$ = temperature gradient in the frozen zone
- $\rho i$ = density of ice
- L = latent heat of fusion
- Vi = volumetric ice flux

The second term of Equation 2.10 represents the heat flux due to the movement of pore ice through the soil. In his final expression, Miller related the rate of heave, frost penetration rate and the velocity of the ice in the frozen zone by:

$$\rho\left[\partial V(\Psi)/\partial z\right]t = \left[\rho i Vh + (\rho-\rho i)V(\theta i)\right]\left[\partial\theta(\Psi)/\partial z\right]t \quad (2.11)$$

where: $\partial V(\Psi)/\partial z$ = change in velocity of water intake with depth
$z$ = depth
$Vh$ = rate of heave
$V(\theta i)$ = velocity of ice content
$\partial\theta(\Psi)/\partial z$ = change in Wu with depth
$\rho$ = density
$\Psi$ = variable to denote amount of ice in pores
$\theta$ = water (or ice) content

Equation 2.11 contains many parameters which are difficult to analyze. Reasonable values for the velocity of ice content and the relation of unfrozen water content with depth would be difficult, at best, to obtain. Still, the expression does couple the effect of moisture migration behind the frost front with the migration of water to the frost front. Miller conceded that the relationship has not yet been experimentally tested. He suggested that Equation 2.11 would be amenable to computer simulation. This would allow comparison with experimental results from future tests.

It is difficult to ascertain the importance of the

movement of pore ice to the overall heave rate from Equation 2.11. Experimental results conducted by this author (see Chapters IV and V) indicate that the net effect of secondary frost heave for engineering problems may not be significant when compared to that of primary heave. This point will be discussed in more detail in following chapters.

## 2.5 Conclusion

In light of the above literature review it becomes apparent that, although it is generally agreed that water can be transported through frozen soil, the mechanisms involved, the rate of migration and the extent of migration are still not adequately defined.

A few points regarding the processes of moisture migration in frozen soil can be synthesized from the literature:

1.) Moisture migration through frozen soil is basically caused by the development of a temperature-induced potential gradient within the unfrozen water film.

2.) The effect of pressure on the system is to increase the film thickness (and therefore, potential) causing a reduction in the suction pressure. The existence of a shut-off pressure in a soil system is questionable

but the available information is not adequate to discount it totally.

3.) The effect of solutes in the soil is to increase the unfrozen water content for a given temperature. This, in turn, reduces the soil's potential for drawing in water. The effects of valence of cations is not totally understood but apparently lower valence cations increase the amount of unfrozen water more than higher valence cations at warmer (near 0°C) temperatures. The effect is not so pronounced at colder temperatures (Nersesova and Tsytovich, 1963).

4.) The rate of moisture migration in frozen soil appears to be dependent upon the temperature gradient in the frozen zone only for steady state frost heave. Although Penner and Ueda (1978) seem to disagree with this statement, further discussion in Chapter V will provide more insight into the relation of temperature gradient with moisture migration.

5.) A regelation process of ice transport in the frozen zone is a possible explanation for the redistribution of water content with time. The connection and interrelationship of regelation with the other processes of moisture transport (such as migration through the unfrozen films) will be discussed in Chapter V).

# CHAPTER III

## Laboratory Testing Program

### 3.1 Testing Program

The objective of the program was to investigate moisture migration and redistribution in a completely frozen silt sample as a result of the application of a temperature gradient through the sample.

The testing program consisted of three closed system and nine open system frost heave experiments. Pressures of 50 and 75 kPa were applied in two tests (A-10 and A-11, respectively) using a weight and hanger system in order to evaluate the effects of overburden pressure on migration characteristics in frozen soil. Except for the load cap (9 kPa), no loads were applied in the remaining tests. A summary of the salient features of each test is presented at the end of the chapter in Table 3.2.

Tritiated water was used in the water supply in several tests as a tracer to help identify where the ice is accumulating. The procedure and results of these investigations are discussed in Section 3.6.

### 3.2 Materials

All tests were carried out using Devon silt. The grain size curve and Atterberg Limits are shown in Table 3.1 and

Figure 3.1 for two samples. Sample A was used in the first six tests while Sample B was used in the remaining six. The slight difference in properties is attributed to the fact that the samples were obtained from different strata. It is believed that this small variance in soil type should not affect test results to any noticeable degree since many other variables in the testing procedure exist that dominate heave and migration characteristics. These variables are explained further in the following chapters.

TABLE 3.1

Index Properties of Devon Silt

| Property | A | B |
|---|---|---|
| Liquid Limit (%) | 41.6 | 40.4 |
| Plastic Limit (%) | 21.7 | 19.7 |
| Plasticity Index (%) | 19.9 | 20.7 |
| Specific Gravity | 2.7 | 2.7 |
| Clay Sizes (%) | 30.0 | 25.0 |

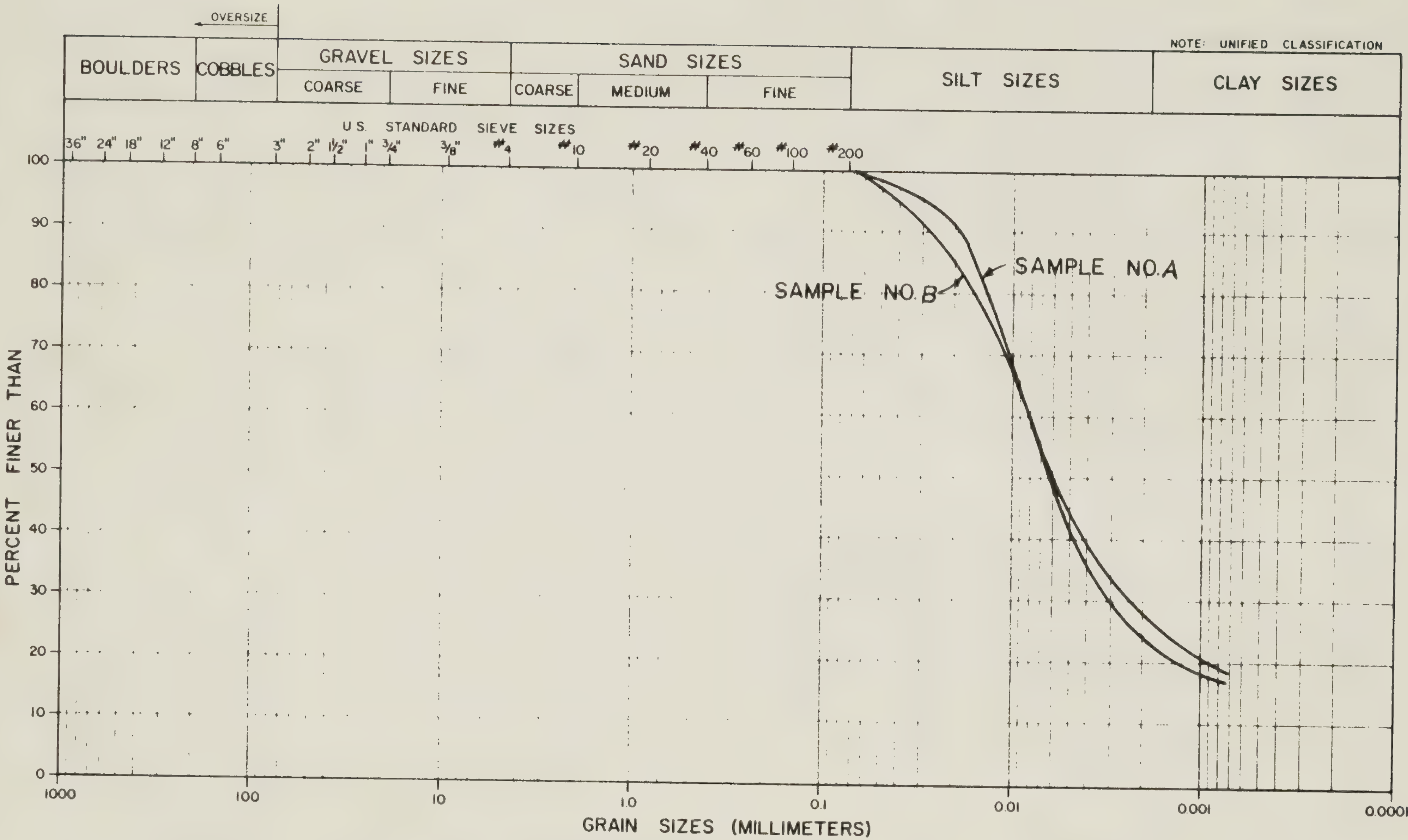



FIGURE 3.1 : GRAIN SIZE DISTRIBUTION CURVES FOR DEVON SILT

## 3.3 Equipment

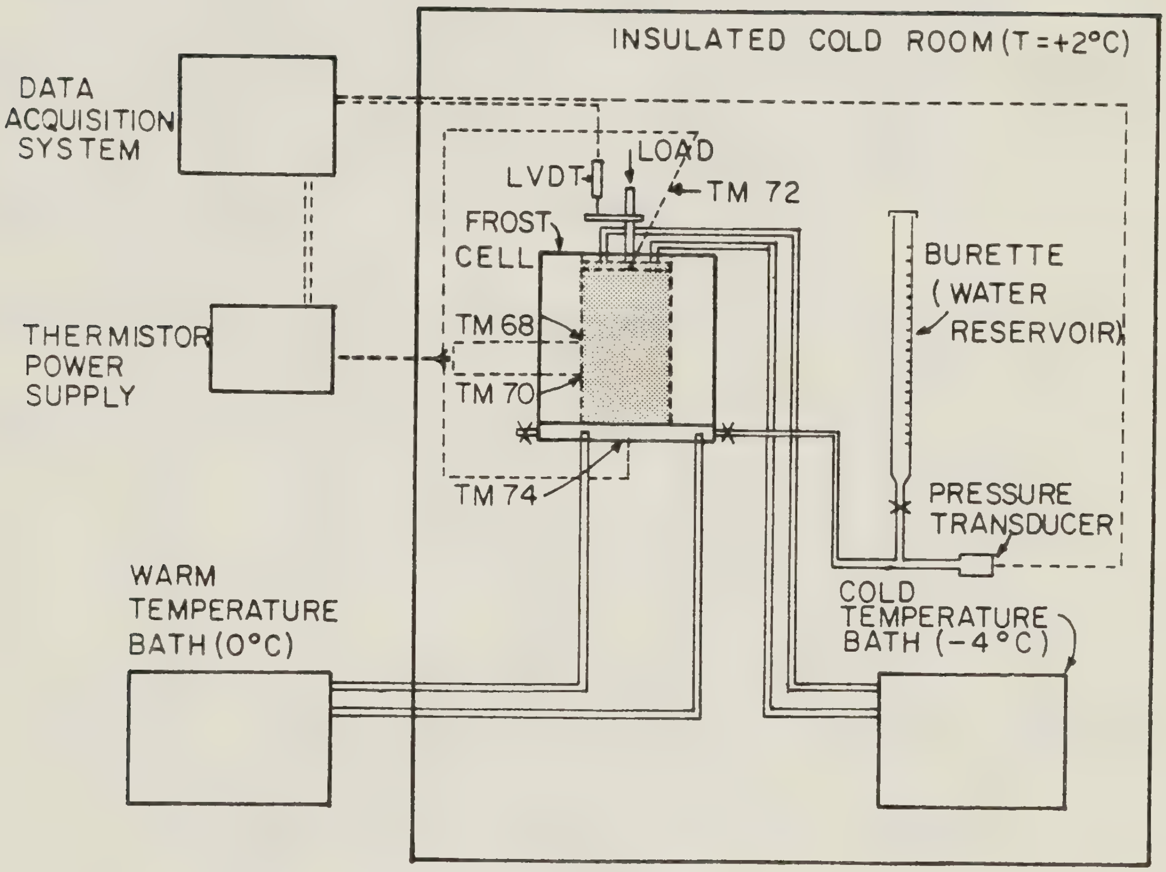
A schematic diagram of the experimental set-up is given in Figure 3.2.

The cold room was maintained at a fairly constant temperature of 2°C. However, because defrosting was necessary for long term use to prevent ice build-up on the refrigeration pipes, the room was heated to at least 10°C. for approximately one half hour each day.

A frost cell was the primary experimental apparatus used in the test program. The frost cell is an adaptation of the permode (Permafrost Oedometer; see Roggensack, 1976 and Hill, 1977). Details of the construction are given in Figures 3.2, 3.3, 3.4, 3.5, and 3.6.

The frost cell is a 10 cm. I.D. teflon lined cylinder. The outer jacket is machined P.V.C. pipe which acts somewhat as an insulator and also provides lateral restraint during application of a load. A styrofoam cylinder was fitted around the P.V.C. and sealed with a silicone sealant to provide additional insulation.

The top piston was used as a heat sink through which anti-freeze from the cold temperature bath was pumped. It also distributed the load, which was applied by means of a hanger-weight assembly. Double "O" rings set in grooves provided an adequate seal. No moisture transfer was allowed




TM............ THERMISTOR
X.................. VALVE
----------- ELECTRONIC LEADS
SAMPLE

FIGURE 3.2 : SCHEMATIC DIAGRAM OF EXPERIMENTAL SET-UP.

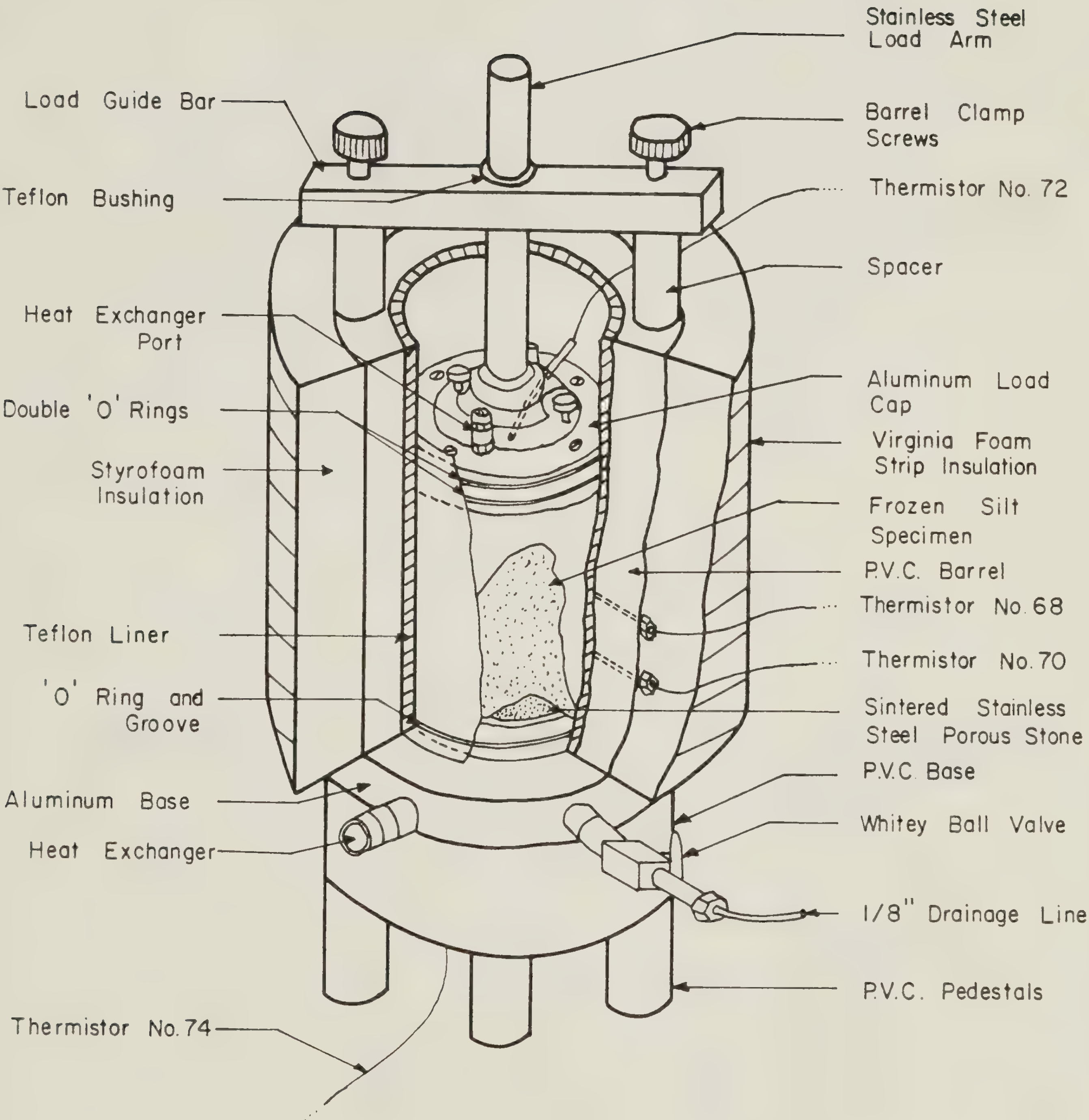



FIGURE 3.3 CUTAWAY VIEW OF FROST CELL

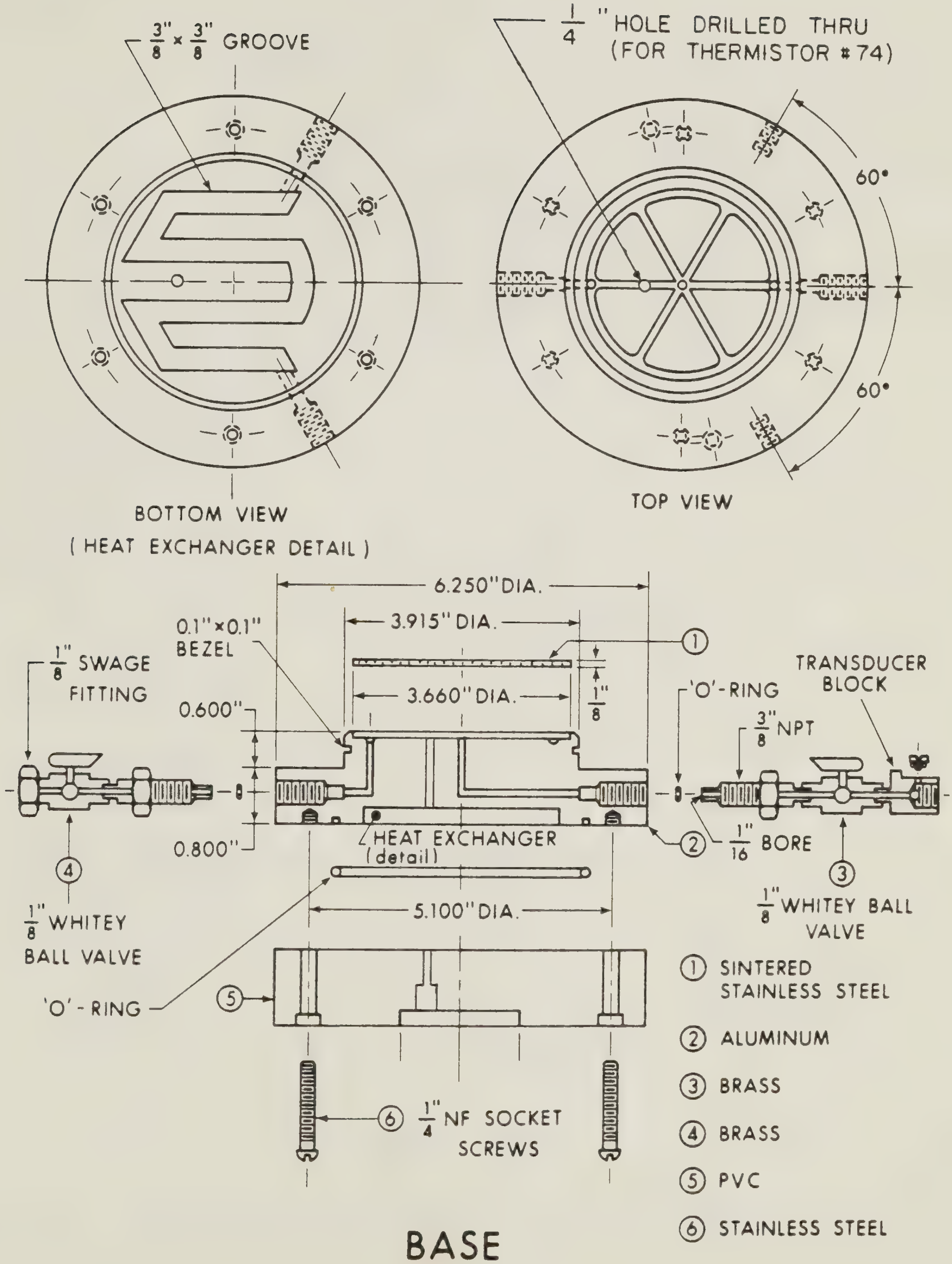
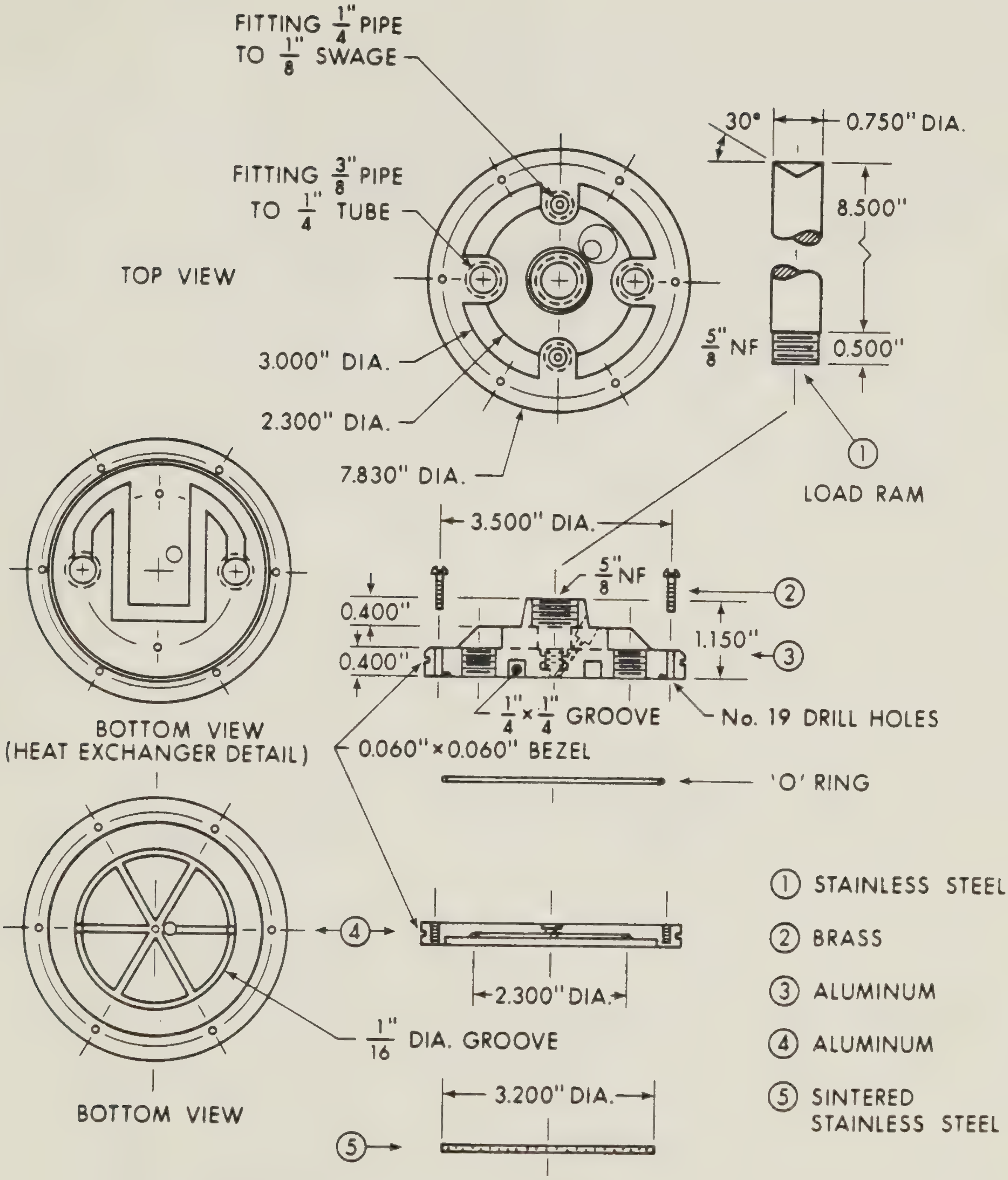



FIGURE 3.4 : DRAWING OF BASE




LOAD CAP

FIGURE 3.5 : DRAWING OF LOAD CAP

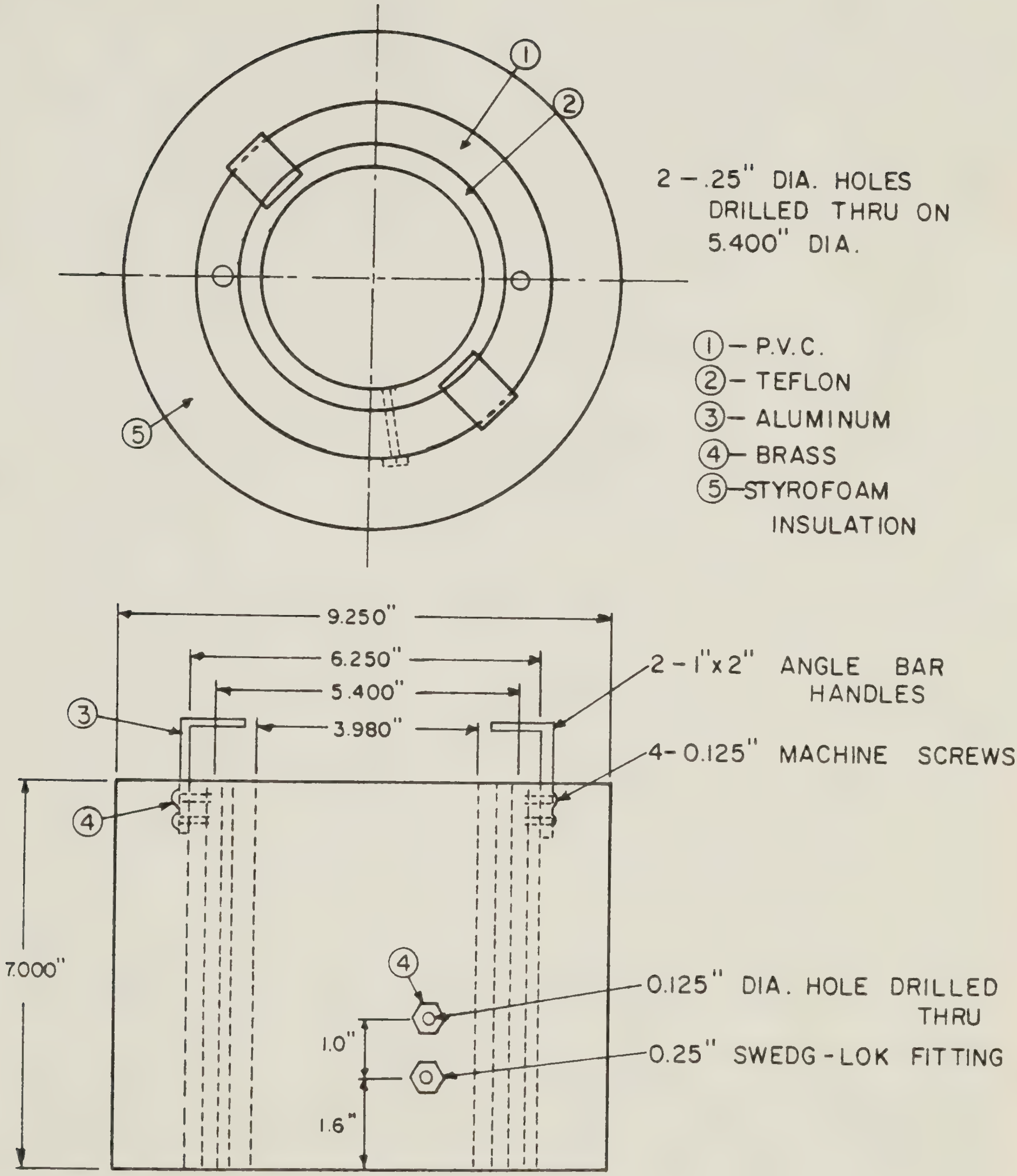



FIGURE 3.6: FROST CELL BARREL

from the top of the sample. The base plate also served as a heat sink through which antifreeze from the warm temperature bath circulated. The external source of water from the burette was made available to the frozen sample at the bottom through a sintered steel porous stone set into the base. For closed system tests the porous stone was replaced with a copper disc. The seal at the bottom was accomplished with a thick "O" ring set in a groove.

Thermistors (described below) were encased in 1/8 inch O.D. hollow steel tubing which was fitted with a Swedg-Lok fitting. This enabled easy replacement of thermistors if a malfunction should occur. Thermistors were installed in the top and bottom plates just under the porous stone or copper plate. Two thermistors were also installed in the side walls in order to measure the temperature distribution along the sample. They were spaced 2.5 cm and 5.0 cm from the base. The configuration is shown in Figure B-1 in Appendix B.

The temperature at the top (warm side) and bottom (cold side) was controlled by continuous circulation of anti-freeze/water mixture. The fluid temperature was maintained by separate Hotpack constant temperature baths. Temperature fluctuation was normally under 0.05°C over several weeks. The rate of pumping was 80 ml/sec.

A sensitive differential pressure transducer (0-4 kPa; 0-0.5 psi) was joined to the water supply (burette) via a

"T" connection. The output (in volts) from the pressure transducer was then connected to the data acquisition system described below. This set-up, shown in Figure 3.7, allowed continuous reading of water height in the burette (and, therefore, rate of water intake into the sample) over extended periods of time obviationg manual recording of information . Checks were made periodically and the system proved very reliable producing results within 2% of the actual readings (which could be interpolated to 0.1 ml).

Vertical displacements were monitored using a 6 volt excitation LVDT (Linear Voltage Displacement Transducer) accurate to 0.025 cm. The output (in volts) was recorded on the data acquisition system.

It was decided that, due to their increased accuracy and adaptability to the data acquisition system, Atkins No. 3 thermistors would be used instead of thermocouples. The thermistors were calibrated with a Hewlett-Packard quartz thermometer accurate to 0.001°C. However, due to the inherent difficulties in calibration and to changes in thermoconductivities of the calibration medium (ethyl alcohol-water mixture) and the material into which the thermistors were placed (aluminum, pvc, teflon) it is expected that the temperature measuring system is accurate only to 0.05°C.

A Fluke data acquisition system employing a digital

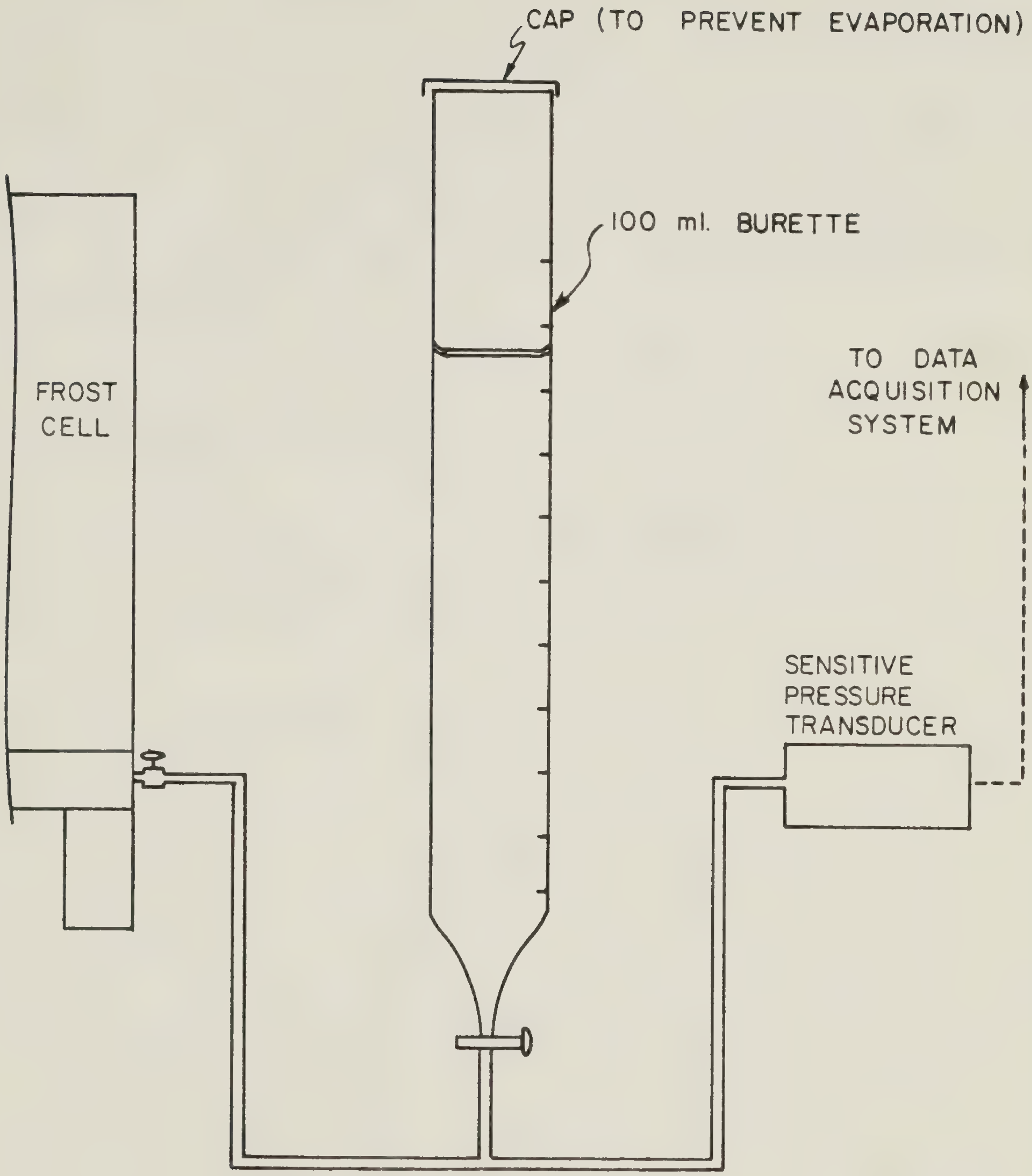



FIGURE 3.7 : SCHEMATIC DRAWING OF AUTOMATIC WATER-INTAKE MEASURING SYSTEM.

voltmeter was used to obtain output from thermistors, LVDT and pressure transducer. The information, in volts, was stored on a cassette tape by means of a Techtran cassette recording device. The cassette tapes were periodically "dumped" into a computer file space where the output could easily be converted and plotted using relatively simple computer programs.

### 3.4 Sample Preparation

A description of sample preparation is best accomplished by dividing the process into four steps:

1.) Slurry Preparation was achieved by mixing air-dried silt with de-aired, distilled water. The mixture was allowed to stand overnight to permit saturation. The slurry was then poured into a vacuum dessicator. A vacuum of approximately 680 mm Hg was applied while vigorously vibrating on a shaking table for at least 40 minutes. This process removed nearly all entrapped air as was demonstrated by Hill (1977). After preparing samples in the same manner described above Hill performed B tests which resulted in values of B very near unity signifying a high degree of saturation. Water contents of the slurry ranged between 50% and 60%.

2.) <u>Consolidation</u> of the silt slurry to 100 kPa was performed in two stages in an oversized (12.3 cm, I.D.) consolidometer. The time to t-100 as determined from the change in height - log time plot averaged near 18 hours for a sample with an initial height of 18 cm (Final height = 12 cm). The water content after consolidation averaged between 30 and 33 percent and was found to be relatively uniform throughout the sample.

3.) <u>Initial Freezing</u> was accomplished by slowly immersing the sample into an open container filled with liquid nitrogen (T=-195°C = -319°F). The purpose of using liquid nitrogen was to freeze the sample fast enough to preclude moisture redistribution within the sample. As it turned out, although small in magnitude, some redistribution of moisture to the colder side did occur in most of the samples. Figures 3.8 and 3.9 show typical water content profiles before and after freezing. Note in both figures the drop off in water content near the top. This is due to the fact that the freezing process (in liquid nitrogen) was terminated before complete freezing was accomplished. The sample was stored in a cold room (T = -10°C) and freezing continued at a much slower rate, thereby allowing time for some moisture to migrate from the top portion to the much colder lower portion of the sample. When the

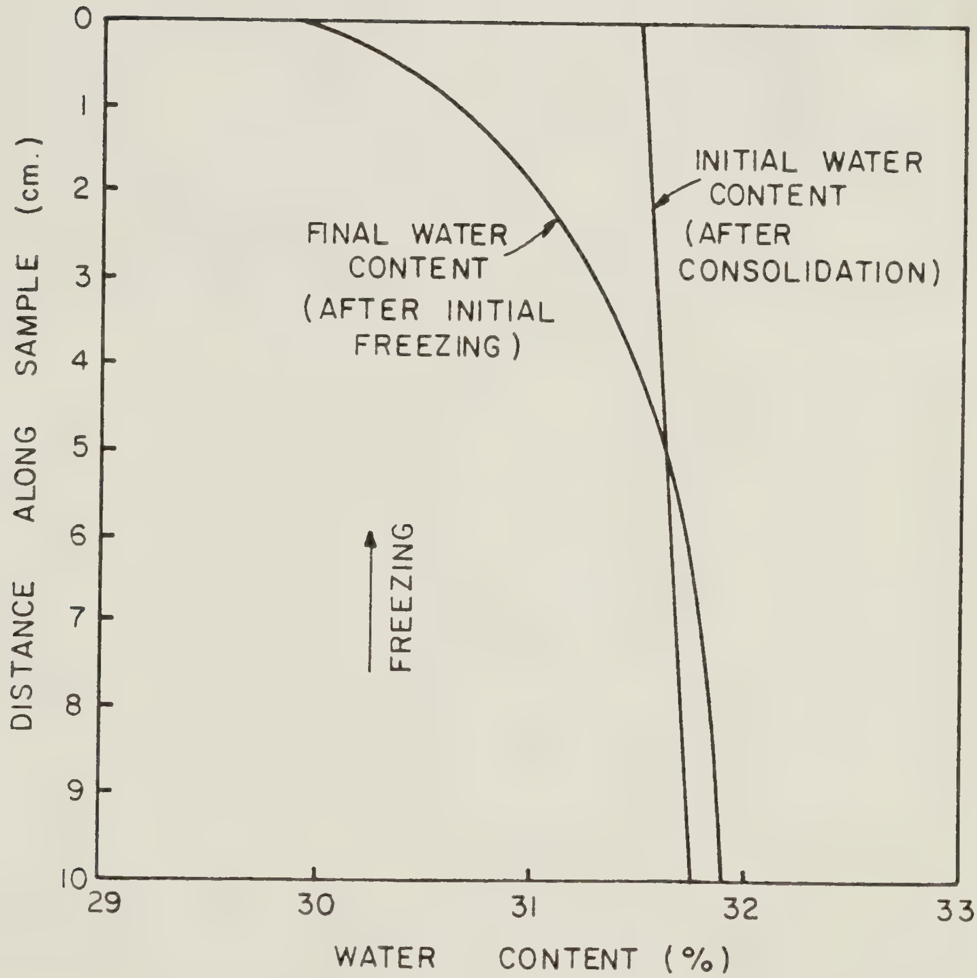



FIGURE 3.8: MOISTURE REDISTRIBUTION DUE TO INITIAL FREEZING.

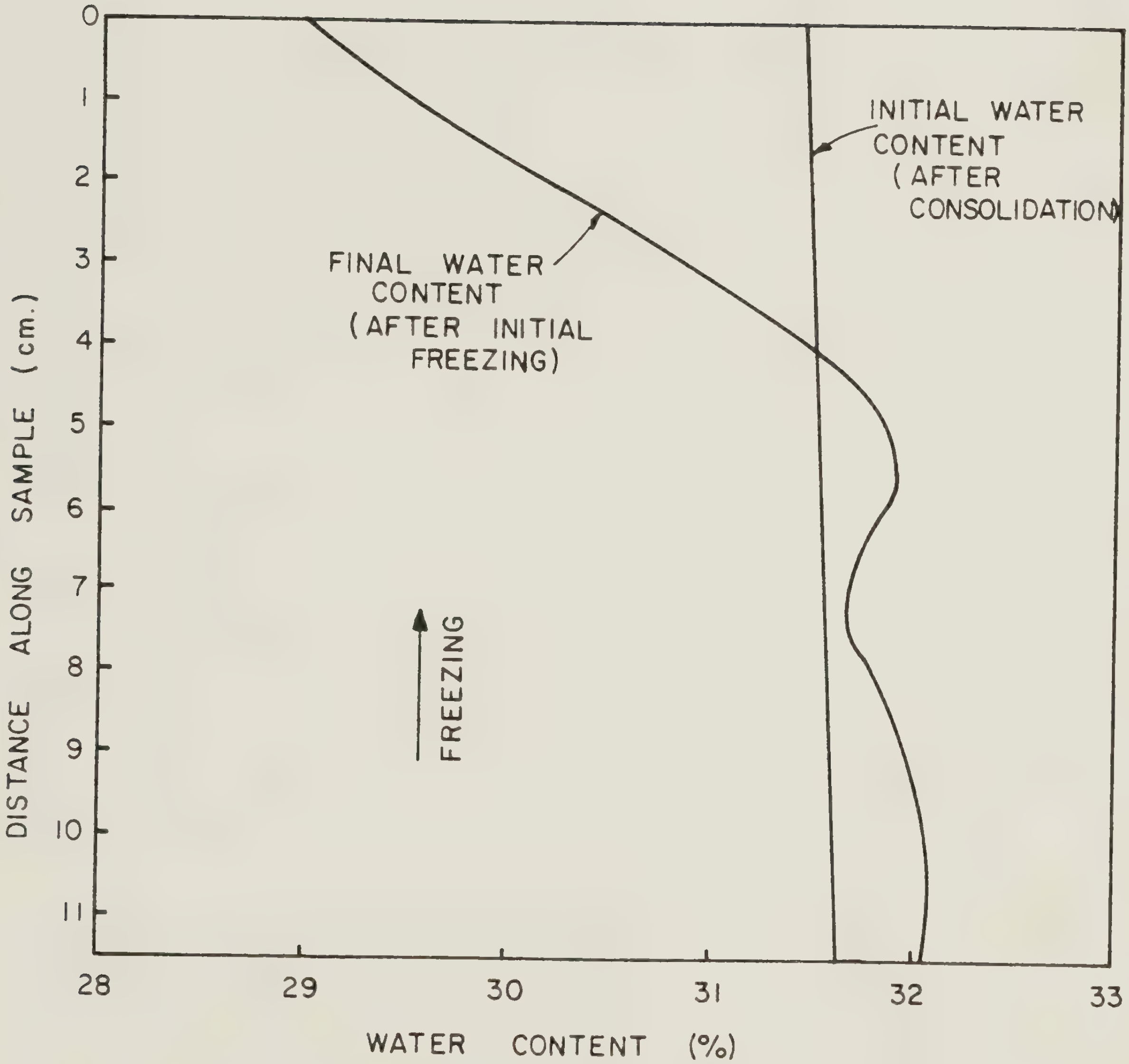



FIGURE 3.9: MOISTURE REDISTRIBUTION DUE TO INITIAL FREEZING.

sample was trimmed (see step no.4) much of this non-linear portion was cut off at the top and at the bottom. Nevertheless, in most of the tests the water content at the top was usually somewhat less than at the bottom. However, it is expected that the shape of the distribution is consistent throughout the tests. The fact that the distribution is linear in the lower portion of the sample is important in that it is in this section where most of the moisture migration occurs.

Freezing soil in liquid nitrogen is a delicate process. If the sample is immersed too quickly very high stresses develop causing cracking to occur. By trial and error it was found that an optimum time to freeze a 12 cm high by 12.4 cm diameter sample was about 10 to 15 minutes. Theoretically, it can be shown that the time for complete freezing should be 23 minutes if the bottom of the sample is kept at the liquid nitrogen surface. However, the sample was immersed very slowly as the soil froze thus increasing the freezing rate.

The frozen sample was allowed to warm to the cold room temperature of -10°C. The ends of the sample were marked 'top' and 'bottom', respectively, and the entire sample was coated with a layer of ice by spraying with an atomizer, wrapped in plastic and stored in a freezer at T= -15°C.

Hoekstra (1965) froze a rectangular clay sample between

copper plates at a temperature of -30°C. However, his sample was much smaller (2 X 6 X 6 cm) and therefore freezing at this temperature was accomplished fast enough to prevent moisture redistribution. Time to complete freezing for Hoekstra was 18 minutes which is comparable to that found using the method presented above.

4.) Trimming the sample was the final step in sample preparation. The ends of the frozen sample were sawed with a band saw to provide flat surfaces for a proper fit in the lathe. An extra slice at each end was also cut off for the purpose of water content determination. It was assumed that the water content distribution was linear through the sample.

The sample was then trimmed in the lathe to a diameter slightly smaller than the inside diameter of the frost cell (approx. 9.95 cm). The top and bottom surfaces of the sample were milled flat. The sample was weighed, measured and then coated with vacuum grease (sides only) to help prevent any dessication. A thin rubber membrane was fitted around the sample completing the sample preparation.

## 3.5 Test Procedure

### A. Open System Tests

Tests A-5 through A-13 were considered open system tests in that the frozen sample was allowed access to a water supply (from the bottom only) throughout the experiment.

The pretrimmed sample was placed on filter paper which covered the porous stone. Prior to this step the porous stone had been boiled in water for de-airing and, as well, all water lines were de-aired. The rubber membrane was then sealed at the bottom with an O-ring. The top piston was inserted and sealed with two smaller O-rings. The cylinder portion of the frost cell was slipped over the assembly and clamped in place. The anti-freeze lines leading to the top piston were connected and circulation commenced (the lines to the base were previously connected). The LVDT was set in place and all wiring for the LVDT, thermistors and pressure transducer were connected. The burette was filled and the top covered to prevent evaporation.

For those tests (A-6 through A-13) where tritiated water was used as a tracer the only deviation from the above procedure was the replacement of H2O in the burette with H3O water prior to sample installation.

Tests A-8, A-9, A-12 and A-13 were performed with

frozen silt samples that contained a pre-made ice lense. The lense was imbedded at or near the bottom of the sample. After the trimming process was completed a thin slice of the sample was cut off near the bottom and a thin (0.5 - 1.0 cm) pre-made ice lense was frozen in between the two soil cylinders. In Test A-9 the lense was frozen directly on the bottom of the sample. Care was taken to prevent any air voids from forming between the ice-soil contacts. Gaps due to irregularities on the sides of the ice lense were filled in with water which quickly turned to ice. The sample was finally retrimmed by scraping to remove any extra ice on the sides. Testing proceeded as before.

Two tests, A-10 and A-11 involved the application of loads during testing. A pressure of 100 kPa was applied at the start of A-10. After almost one day no heave was noticed and 50 kPa was removed. Water intake commenced immediately. A pressure of 75 kPa was applied from the beginning of Test A-11. Re-consolidation near the bottom due to partial thawing delayed water intake for 7-9 hours. Open system tests were usually conducted for a period of 5 to 6 days; however, Test A-13 lasted a total of 22 days. Cold temperatures (top) remained constant at around -4.2°C. Warm temperatures (bottom) were relatively constant at or slightly above 0°C. Upon completion of the tests samples were again weighed and measured. Photographs were taken to record the position of the ice lense.

The sample was then cut into slices approximately one centimetre in thickness. The outside portion of the slices were trimmed and the remaining portion weighed and oven-dried (105°C) to determine the water content. In tests where tritiated water was used, approximately one half of each slice thought to contain tritium was stored in a water-tight container. These samples were later tested for tritium concentrations as described in Section 3.6 on tritium analysis.

B. <u>Closed System Tests</u>

Three tests, A-2, A-3 and A-4 were conducted as closed system tests in that the sample was not allowed access to water throughout the experiment.

Prior to testing the porous stone in the base was replaced by a round copper plate sealed with vacuum grease in the same manner as the top plate had been.

The test procedure involved only the placing of the sample in the frost cell as described for the open system tests. No monitoring of vertical deformations was required but an LVDT was nevertheless installed as a check. As expected, no observable deformations were recorded. Temperatures at the top and bottom plates were the same as for open system tests, -4.2°C and 0°C, respectively. Duration of the tests ranged from 11 to 16 days.

Upon completion of the tests, samples were cut into slices for water content determination as previously described.

## 3.6 Tritiated Water Analysis

### 3.6.1 Background

The use of tritiated water (generally denoted as H3O) to trace the movement of water through soil is well documented (Guidebook on Nuclear Techniques in Hydrology, 1968). One important application is the monitoring of groundwater infiltration rates in order to evaluate the effect on apparent groundwater recharge (Blume, et al, 1967).

Tritium is essentially a water molecule with an atomic weight of 3 in which the two hydrogen atoms each contain two added neutrons (in addition to the one proton). Being unstable, the isotope decays by emitting very small beta-particles. The half-life of tritium is about 12.5 years. Due to the fact that only a small concentration is required for detection (less than 1 μCi (micro curry) per liter of H2O) and that the size of the emitted particles is small and not very strong radioactively, the use of tritiated water in the laboratory is relatively harmless. Beta-particles cannot penetrate the skin but precautions should be taken to avoid

inhaling vapors directly.

The main advantage of using tritiated water as a tracer is that it causes no change in the natural condition of the soil (Corey and Horton, 1968). The only difference physically between H30 and H20 is that the isotope has a slightly higher density due to the added neutrons, but the difference is so slight as to be insignificant. If a soil is completely homogeneous and isotropic with respect to density, material type, pore size, permeability, etc., it would be expected that a layer of tritiated water would move through the soil along with and at the same rate as the adjacent ordinary water. However, in studies where tritiated water was injected into a natural soil system in the field (Zimmerman, et al, 1968) it was noticed that, in time, the originally distinct tagged layer would become more diffuse indicating some dispersion of tritiated water ahead of the tritiated/non-tritiated boundary. This is due primarily to the variability of soil permeability at the microscopic level. Water will travel faster through the larger voids; therefore a "fingering" effect occurs in these areas whereby some tritiated water will move ahead of the interphase and disperse with the ordinary water until ultimately a distribution of H30 concentration results rather than a distinct layer. Depending upon the nature of the distribution one can evaluate the extent of migration (and therefore velocity) of the tritium by choosing a zone at

which the distribution is greatest.

### 3.6.2 Procedure

The test set-up was identical to that described in Section 3.5 except the distilled, de-aired water in the burette was replaced by distilled, de-aired tritiated water. The concentration of tritium was about one micro curry per 500ml H2O. Also, the porous stone was boiled in the tritiated water to ensure that any water entering the sample had the same concentration of H3O. The frozen sample was allowed access to the tritiated water at the start of the test and migration into the frozen soil commenced. The sample was cut into thin slices, approximately one centimetre in thickness, after the test was completed in the same manner as in non-tritiated tests. The outside portion was trimmed off to eliminate variations due to boundary effects. Approximately one half of each slice was then placed in air tight containers and stored. The samples were allowed to thaw.

The first step in the analysis of tritium concentrations was the extraction of water from the soil. After thawing, an equal amount of soil (by weight, approximately 20 to 30 grams) from each slice was put into separate test tubes. The tubes were placed in a centrifuge for 10 minutes at 17,000 rpm. One gram of clear, tritiated

water was then extracted and stored in airtight containers. A standard liquid scintillation process was performed on the liquid in the viles. The results of this process are expressed as H30 disintegrations per minute (DPM). For a given amount of tritium contained in the extracted liquid there exists a certain DPM. For easier interpretation, the DPM count of water from each slice is expressed as a percentage of the DPM count of water from the resevoir. For instance, if the water in the layer of soil nearest the porous stone (bottom slice) had been completely replaced with tritiated water the corresponding percentage of new (tritiated) water would be 100%.

A discussion of results from three typical tests is given in Chapter V.

TABLE 3.2

Summary of Tests

| Test No. | Type of Test | Overburden Pressure (kPa) | Pre-made Ice Lense | Tritiated Water | Length of Test (Days) |
|---|---|---|---|---|---|
| A-2 | Closed | 0 | No | No | 16.2 |
| A-3 | Closed | 0 | No | No | 11.8 |
| A-4 | Closed | 0 | No | No | 11.0 |
| A-5 | Open | 0 | No | No | 4.3 |
| A-6 | Open | 0 | No | Yes | 5.0 |
| A-7 | Open | 0 | No | Yes | 5.8 |
| A-8 | Open | 0 | Yes | Yes | 11.0 |
| A-9 | Open | 0 | Yes | Yes | 13.0 |
| A-10 | Open | 50 | No | Yes | 5.7 |
| A-11 | Open | 100 | No | Yes | 5.0 |
| A-12 | Open | 0 | Yes | No | 6.5 |
| A-13 | Open | 0 | Yes | Yes | 22.0 |

# CHAPTER IV

## Test Results

### 4.1 Introduction

A total of twelve tests were conducted on frozen Devon silt. The presentation of the results has been organized into various categories according to the type of test. The three categories and corresponding Section number are given below.

Section 4.2: Closed System Tests

Section 4.3: Open System Tests - No Applied Loads

Section 4.4: Open System Tests - With Applied Loads

Section 4.3 on Open System Tests with no applied loads includes the results from four tests involving the inclusion of a pre-made ice lense in the frozen soil.

Data from each open system test includes the temperature at various points of the sample with time, the change in water intake (or expulsion) with time and the change in height of the sample with time. Data from closed system tests includes only the temperature vs. time relation since no water intake was allowed and an insignificant change in height was observed for these tests. All data is presented in graphical form in Appendix B. The results are arranged in sequential order with respect to test number in order to provide easy reference. A summary of the water

intake and heave rates for the open system tests, as obtained from Figures B-2 through B-13, is provided in Table 4.1 (page 93) along with pertinent details of each test. The numbering and position of each thermistor, as used in the temperature-time plots of Appendix B, is schematically shown in Figure B-1.

The water content profiles, before and after testing, are presented graphically for each test at the end of the appropriate section in this chapter.

## 4.2 Closed System Tests

Three closed system experiments were conducted on initially uniform, frozen silt, A-2, A-3, A-4. The results are presented graphically in Figures 4.1, 4.2, 4.3, respectively.

The results clearly indicate some moisture redistribution in the frozen zone as a result of a temperature gradient of about 0.45°C/cm. The maximum increase in moisture content was about 3% in Test A-2 (time=16 days) and 1 to 1.5% in Test A-3 and A-4 (time=11 days). Note that the maximum extent of redistribution in all three tests occurred in the zone of the sample at which the steady state temperature ranged from -1.5 to -2.0°C.

The initial water contents from the top of samples A-2 and A-4 proved to be lower than expected. As mentioned in

the previous chapter on sample preparation, it was assumed that, in these cases, the low initial water content at the top was due to the removal of the samples from liquid nitrogen before complete freezing had been accomplished. Consequently, the values of the water contents at the top were not indicitive of actual water contents within the sample. A correction, as determined from Figure 3.7, was incorporated into Figures 4.1 and 4.3 in order to establish probable initial conditions.

In all three tests a very thin layer of ice rich soil or pure ice was observed at the top of the sample at the end of testing. Kudryavec, et al (1973) observed a similar phenomenom and explained it as a result of the fact that the top plate acts as an impermeable boundary on which migrating moisture would collect. The mechanism of transfer would probably be vapor diffusion. Due to the high temperature gradients and higher porosity near the top, this explanation seems reasonable.

The temperature control in A-2 (Figure B-2) was not very good, this being one of the first tests. However it is expected that the magnitude of deviation was not sufficient to alter the interpretation of the results to any great degree. The perturbations in the temperatures on the sides of the sample in Tests A-3 and A-4 were due to the 24 hour defrosting of the cold room. Insulation of the frost cell was not sufficient to eliminate the effects of large outside

temperature fluctuations. Since the temperature of the room during defrosting was not consistent over long periods of time, the effects on sample temperature also varied throughout the testing program. In most of the test (open and closed system) some defrost-induced temperature fluctuations were observed. It is thought that, although these perturbations may have some influence on moisture transfer, the effects on the results should be minimal since all moisture contents were obtained from the middle portion of the sample where temperature fluctuations should be insignificant.

Mass balance calculations indicated that approximately 0.015g/hr of moisture was transferred from the warm zone to the colder zones of the sample. The calculations were based on the assumption that no moisture had transferred to the outside. However, in Test A-3 a leak in the membrane allowed some moisture to leave the sample. There was evidence (based on water contents) that moisture accumulation near the outside of the sample was higher than at the center for a given elevation in the sample. This was probably due to the difference in temperature of the center and the outside of the sample. Thus, some moisture migration to the outside must have occurred. Since the rate of moisture transfer was based on water content measurements taken at the center it was reasonable to find that the migration rate in Test A-3 was less than that of the other two tests.

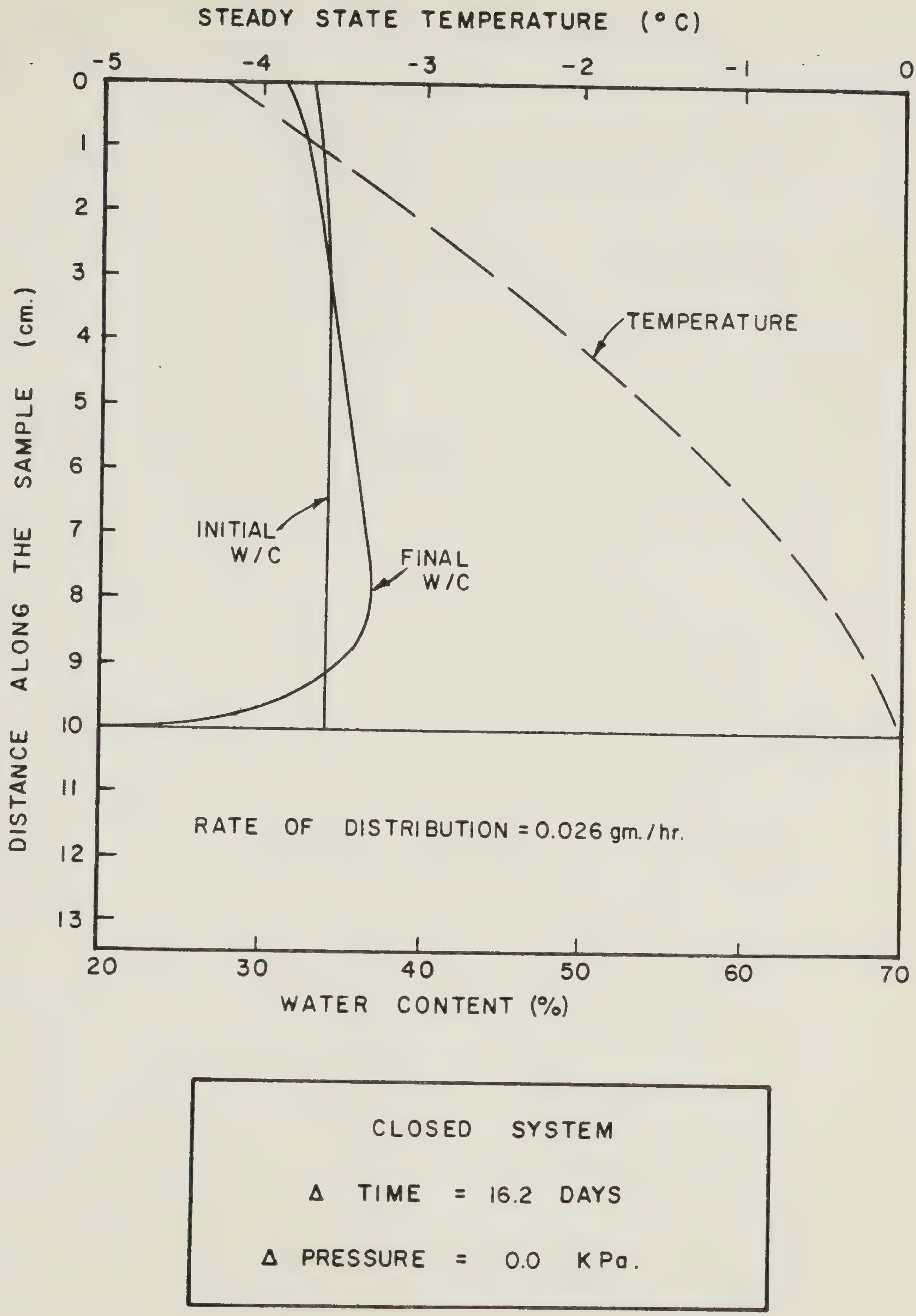
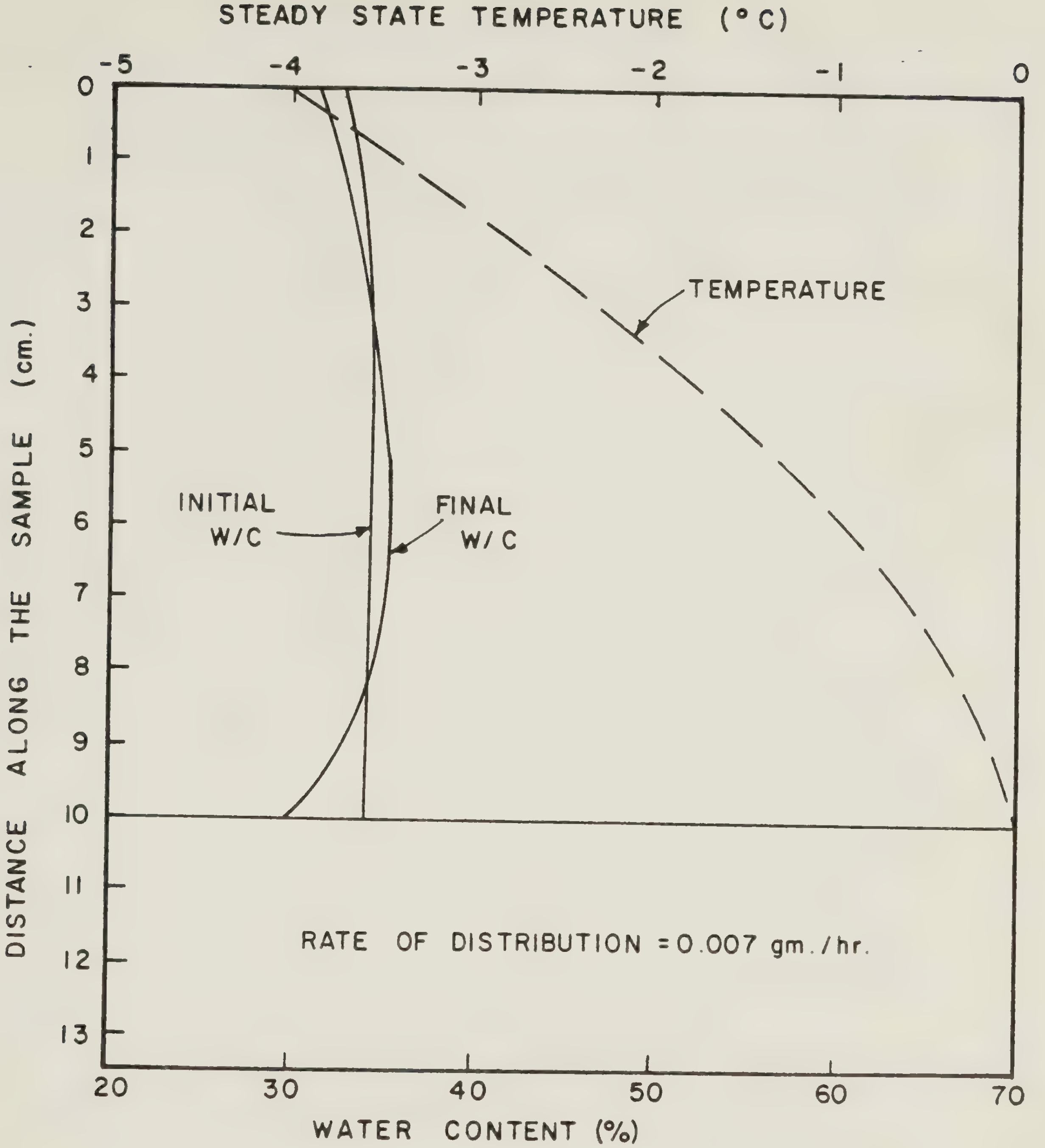



FIGURE 4.1 :MOISTURE MIGRATION GRAPH FOR TEST A-2

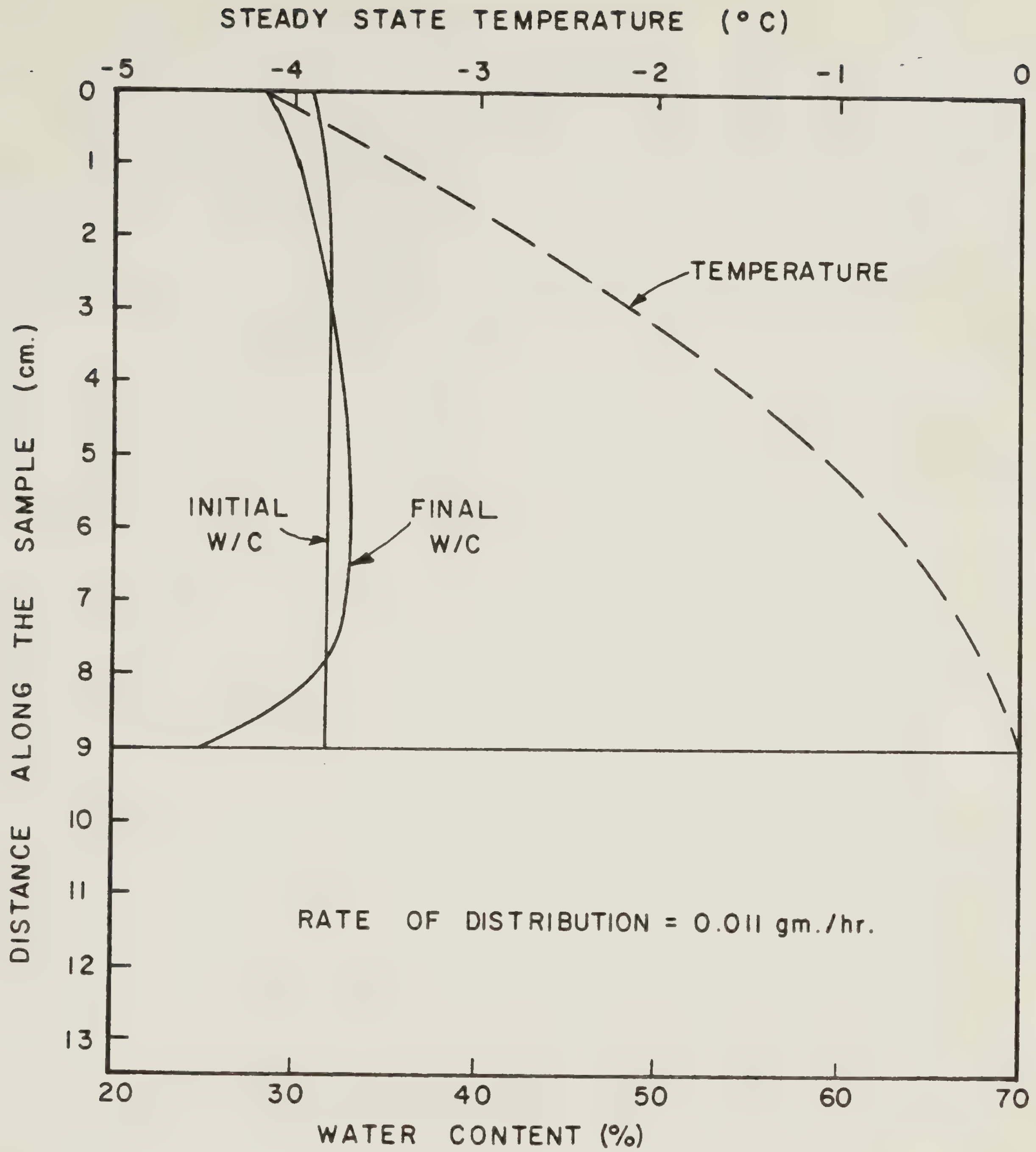



CLOSED SYSTEM

Δ TIME = 11.8 DAYS

Δ PRESSURE = 0.0 KPa.

FIGURE 4.2: MOISTURE MIGRATION GRAPH FOR TEST A-3




CLOSED SYSTEM

Δ TIME = 11.0 DAYS

Δ PRESSURE = 0.0 KPa.

FIGURE 4.3 MOISTURE MIGRATION GRAPH FOR TEST A-4

## 4.3 Open System Tests - No Loads

Seven open system frost heave tests were conducted on frozen silt under a condition of zero overburden pressure. The duration of testing ranged from 5 to 22 days. The results of water content determinations, before and after testing, are given in Figures 4.4 through 4.10. Data concerning sample temperature, heave rates and water intake rates are contained in Appendix B. A summary of frost heave rates and heave characteristics is presented in Table 4.1.

Tests A-5 and A-7 were conducted with samples containing no initial ice lense. The moisture migration graphs (Figures 4.4 and 4.6) show a dramatic increase in water content above the newly formed ice lense. The rate of increase of new water in the frozen zone (above the ice lense only) averaged nearly 0.24 gm/hr. It is not known whether all the water migrated into the frozen soil prior to ice lense formation or whether the migration was an ongoing process occurring simultaneously with ice lense formation.

It should be noted that, although test conditions were kept as consistent as possible, the location of ice lense formation for the two tests was quite different. In Test A-5 the ice lense developed approximately 0.7 cm behind the zero degree isotherm whereas in Test A-7, an ice lense was formed directly on the porous stone. In fact, the filter paper separating the porous stone from the soil was totally

engulfed by the ice lense; thus, no soil at all remained below the lense. It is expected that the interruption in heave and water intake (Figure B-7) near the beginning of the test resulted from the passage of the lense through the filter paper. This anomaly could also be an explanation for the reduction in heave rate (about one half of that for Test A-5) since the suction developed in the more-permeable porous stone should be less than that in the silt.

Test A-6 was originally designed to duplicate Test A-5 in an attempt to achieve reproducibility of results. However, during the first night, the warm temperature bath (thermistor 74) malfunctioned causing the sample to partially thaw from the bottom up. Approximately one half the sample had thawed. As seen in Figure B-6 water intake during the melting period was significant while the amount of heave was minimal. This difference is explained by the fact that the movement of the L.V.D.T. was restricted somewhat for the first 18 hours. The sudden release of the L.V.D.T. arm is seen to have taken place after this time (Figure B-6). It is expected that the actual heave rate was constant from zero to 25 hours. This is shown as a dashed line in the figure. After the warm bath was adjusted down to near 0°C again, refreezing commenced. The heave rate dropped considerably as several thin ice bands developed. Photographs of the sample after testing revealed an ice structure which was basically reticulate in nature with thin

interbedded seams of ice. The highest percentage of ice accumulated near the middle of the sample (Figure 4.5). The water intake and heave rates after freeze-back had been accomplished were greatly reduced to a fraction of those for Tests A-5 and A-7 (see Table 4.1) The important point here is that water was being drawn into the frozen silt at a fairly constant rate resulting in a constant heave rate even though no ice lense had been developed. Frost heave and moisture migration was occurring within frozen soil.

It is noted here that a shift in calibration characteristics of the thermistors had apparently been occurring during the testing program. For tests after A-7 the actual absolute temperature at the four locations of the sample would have been sightly lower than that indicated by thermistor readings. This discrepancy was noticed when the bottom thermistor (No. 74) recorded a temperature of about 0.2°C at a time when ice was forming in the bottom plate and water intake lines. Since recalibration would have been an extremely difficult task it was decided to keep the calibration curves as they were. The temperatures presented at the bottom of Figures B-8 through B-13, therefore, should be approximately 0.2 to 0.3°C lower than shown.

Four open system tests (A-8, A-9, A-12 and A-13) involved the examination of the effects of a pre-made ice lense on the migration characteristics of frozen soil. The duration of the tests ranged from 5.7 to 22 days. The

results, shown in Figures 4.7 through 4.10, reveal that some increase in water content had occurred behind the pre-made ice lense to varying degrees.

The results from Test A-8 were disappointing in that very little migration of water had taken place in the frozen soil. A new ice lense formed near the bottom of the sample and a large increase in water content (final water content = 96%) in the layer of frozen silt between the two lenses had occurred. Relatively no change in the pre-made ice lense was observed and only a slight increase of water content above the pre-made lense was noted.

The phenomenon of the growth of an ice lense directly on the porous stone was examined further in Test A-9. Since the new ice lense in Test A-7 had formed in this manner it was thought that a pre-made ice lense could likewise develop a suction within the ice-porous stone interphase and draw in water to feed the lense. However, the heave and water intake data in Figure B-9A reveal that virtually no heave had eventuated. The slight fluctuations were a result of the opening of the water lines to check whether freezing had blocked the flow of water. Only once had freezing of the lines occurred. The lines were quickly thawed. Since heave could not be induced with an ice lense directly on the porous stone the test was stopped after 70 hours, concluding part A of Test A-9. A thin layer of silt (about 1 to 2 mm thick) was spread on the bottom of the lense.

The second part of Test A-9 (part B) was conducted for another 10 days. No heave or water intake was observed (Figure B-9B). The water intake profile shown in Figure 4.8 reveals some increase in water content above the pre-made ice lense even though no alteration of the lense itself could be identified by visual inspection.

A third open system test with an embedded pre-made ice lense was conducted in an attempt to obtain more information on the migration of moisture from the lense into the frozen soil. It was the original intention to continue Test A-12 for a long period of time under warmer temperatures (cold temperature=-2.5°C) but problems with temperature control of the cold room caused cessation of the test at 5.7 days. Even after this relatively short period of time some increase in water content above the pre-made ice lense had occurred (Figure 4.9).

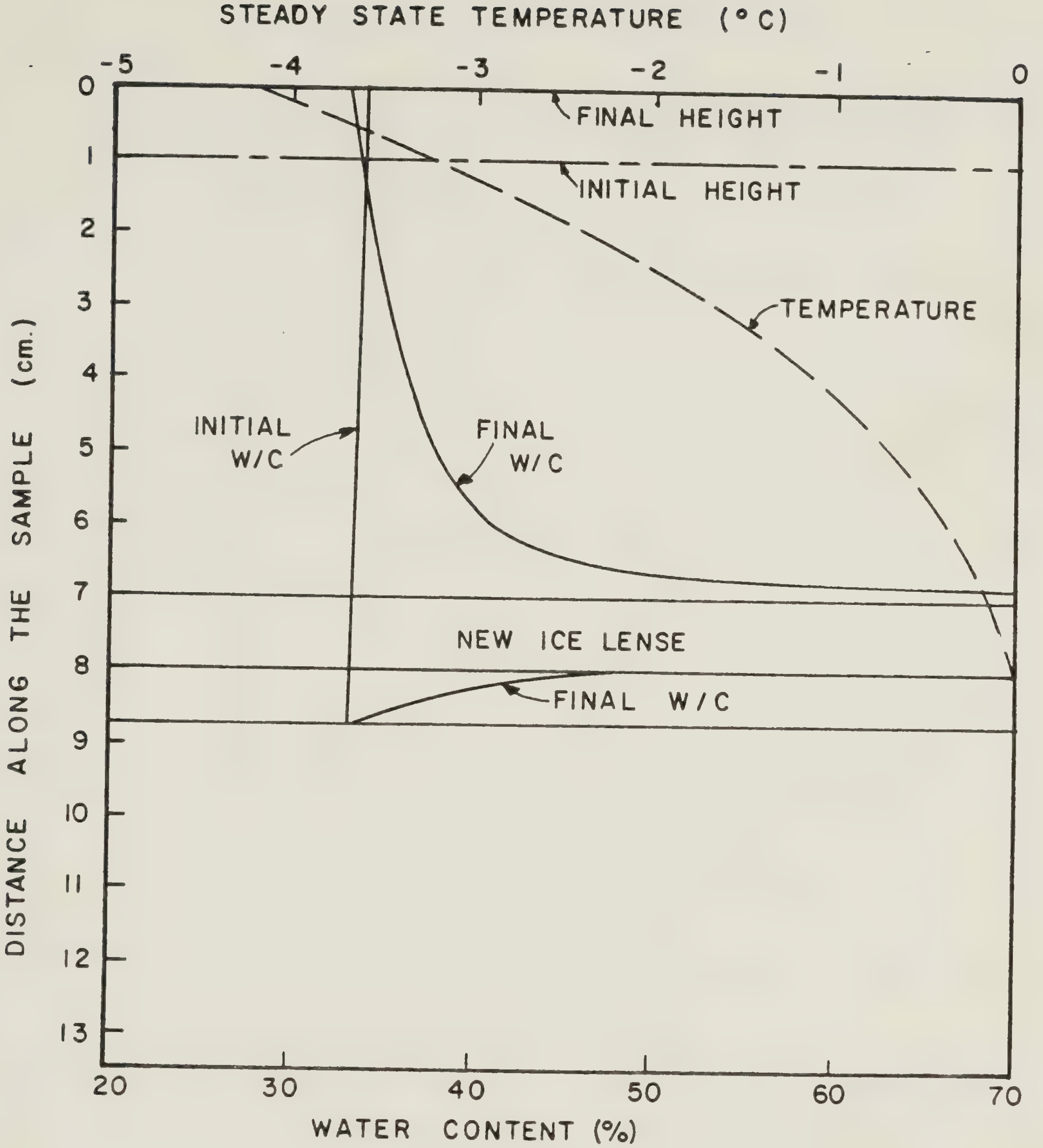
Abrupt changes in heave characteristics were noticed after about 55 hours of testing. The results in Figure B-12 indicate that partial melting from the sides must have resulted when the cold room warmed. Some expulsion of water due to consolidation occurred. The configuration of soil within the freezing zone probably changed since water intake and heave stopped at this point. The lowering of the cold-side temperature to -3.5°C was not effective in inducing heave. Ice formation was reticulate in nature; thin lenses

were on the order of 0.5 mm thick.

Test A-13 was the final experiment involving a pre-made ice lense. The experimental setup was similar to that of Test A-12. The cold temperature, however, was set at -3.7°C to prevent the sample from thawing as happened in Test A-12. The experiment lasted 22 days.

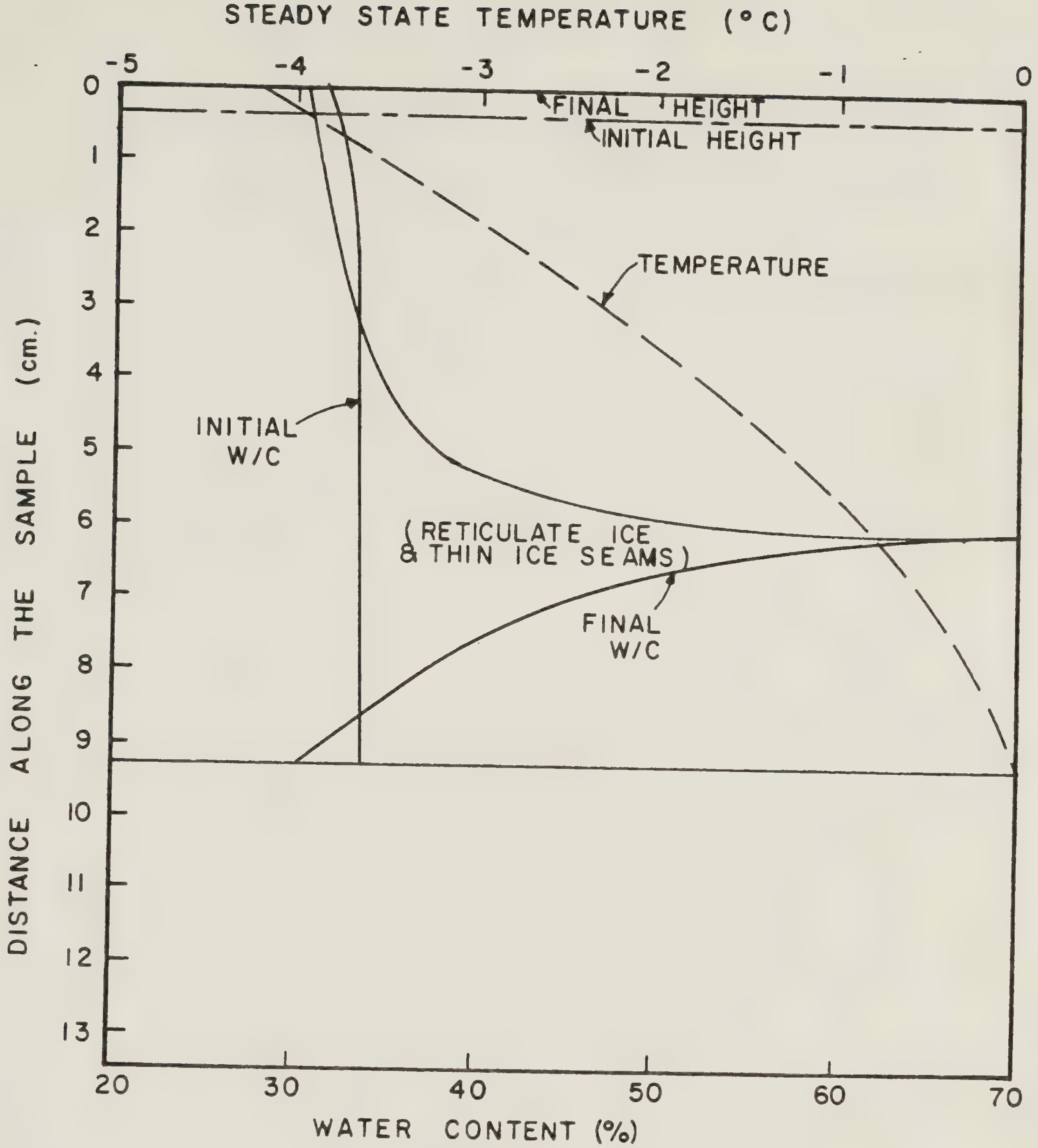
Results from Test A-13 proved to be very interesting. The moisture redistribution curve (Figure 4.10) shows conclusively that moisture from the pre-made ice lense migrated into the frozen soil above the lense. A significant increase in water content was observed in this soil. Moreover, the structure of the pre-made ice lense after testing was observed to have been very pitted and porous in nature within the upper (colder) portion of the lense. This can be seen clearly in Figure 4.11b.

The new ice lense which developed below the pre-made lense is shown in Figure 4.11a. A thin layer of soil separated the new ice lense from the porous stone. The layer of frozen soil between the two ice lenses contained a very high ice content with several thin ice seams interbedded. As with the other open system tests, the heave rate decreased with time. Figure B-13 illustrates that a significant decrease in heave rate occurred after about 100 hours. However, no noticeable change in temperature or other boundary conditions was observed.



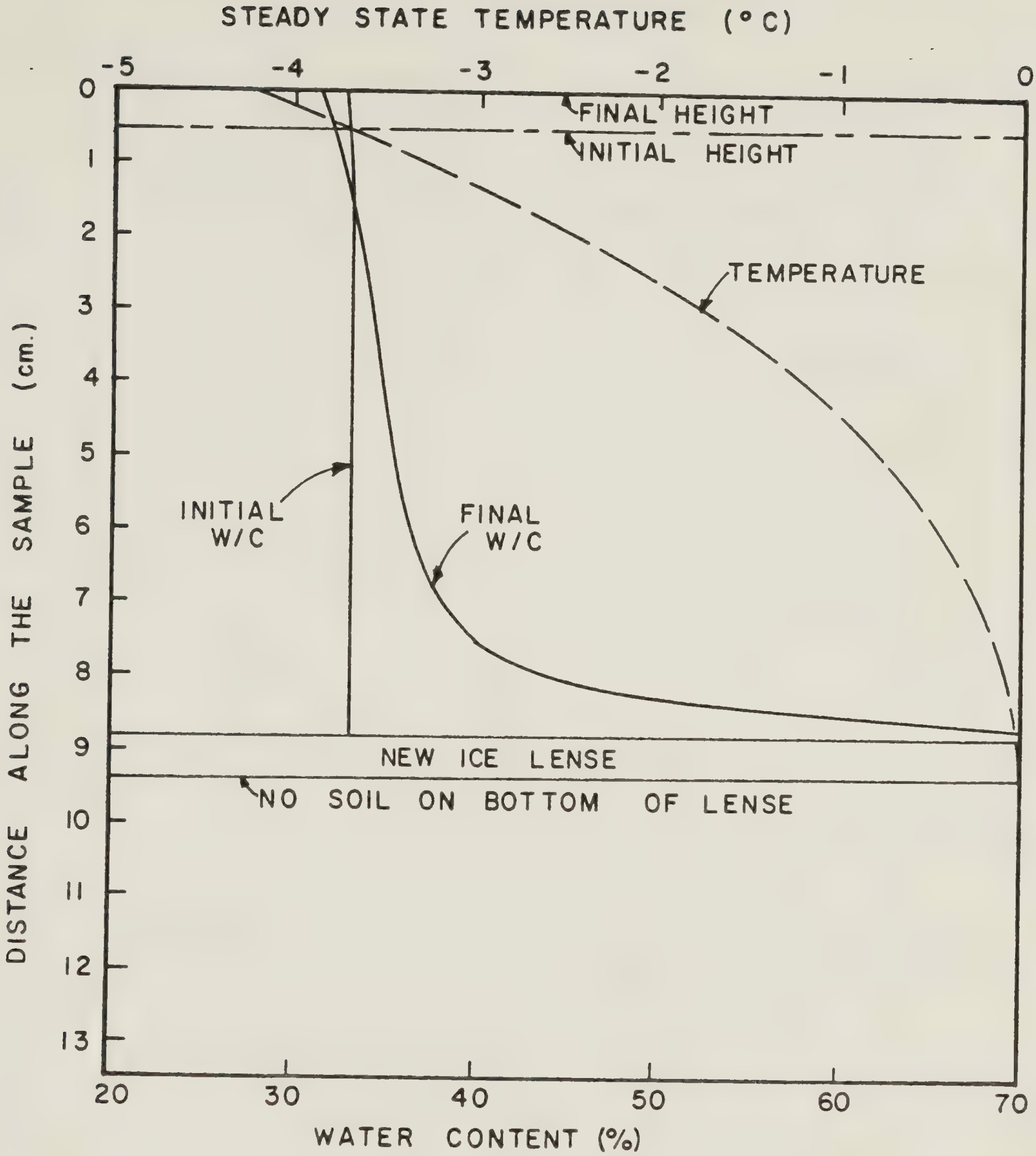

| |
|---|
| OPEN SYSTEM |
| Δ TIME = 4.3 DAYS |
| Δ PRESSURE = 0.0 KPa. |

FIGURE 4.4: MOISTURE MIGRATION GRAPH FOR TEST A-5




| OPEN SYSTEM |
|---|
| Δ TIME = 5.0 DAYS |
| Δ PRESSURE = 0.0 KPa. |

FIGURE 4.5 MOISTURE MIGRATION GRAPH FOR TEST A-6



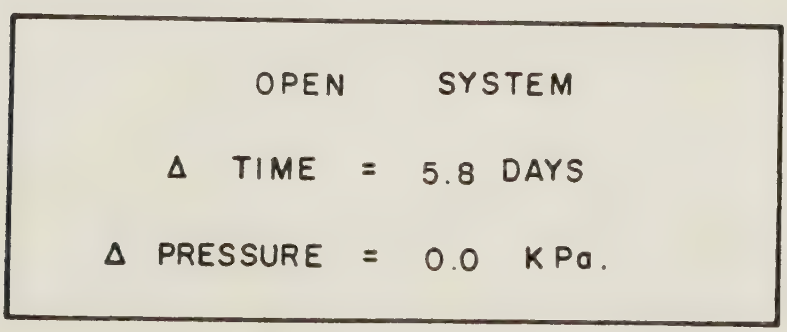
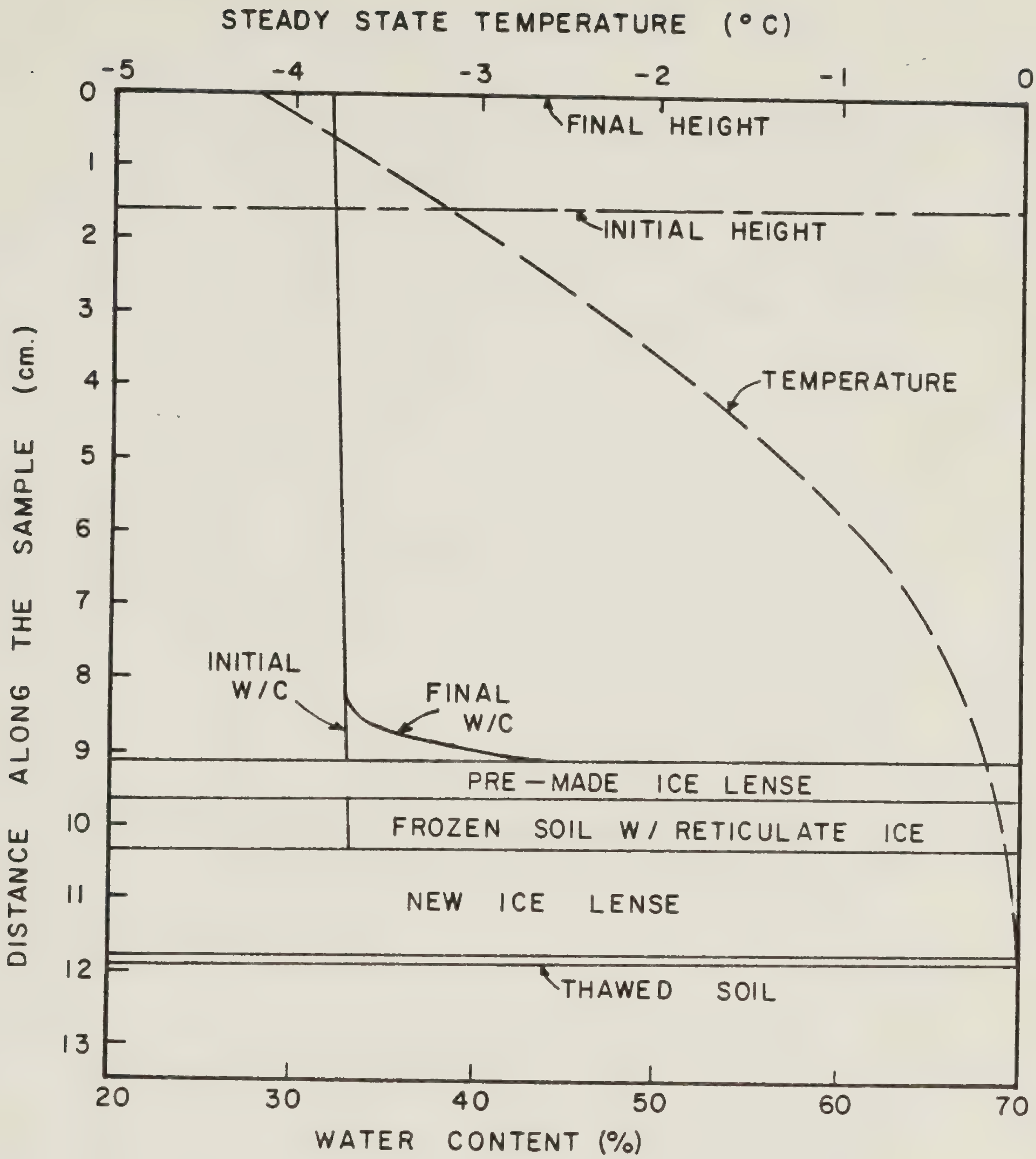




FIGURE 4.6: MOISTURE MIGRATION GRAPH FOR TEST A-7




| OPEN SYSTEM |
| --- |
| Δ TIME = 11.0 DAYS |
| Δ PRESSURE = 0.0 KPa. |

FIGURE 4.7: MOISTURE MIGRATION GRAPH FOR TEST A-8

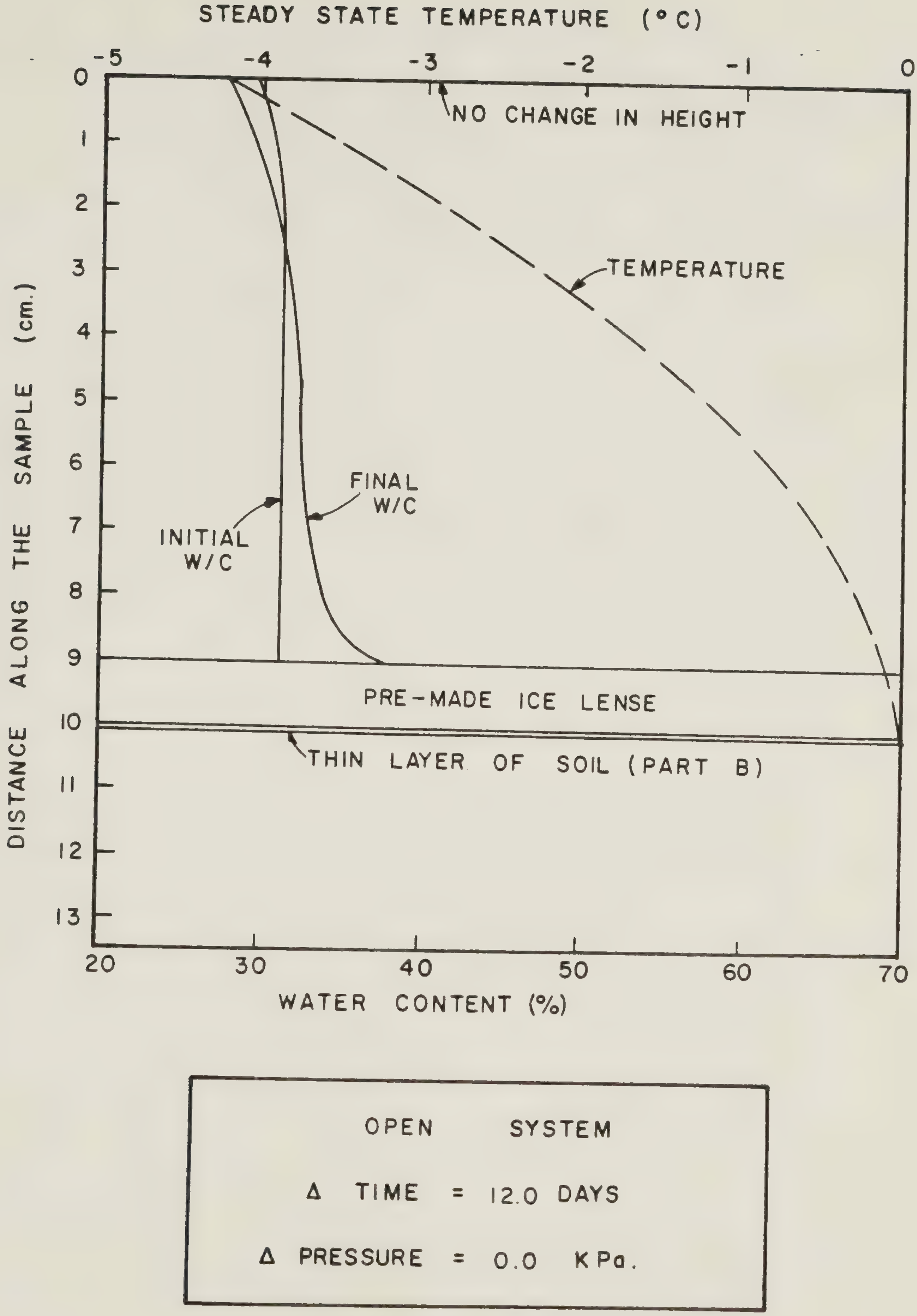
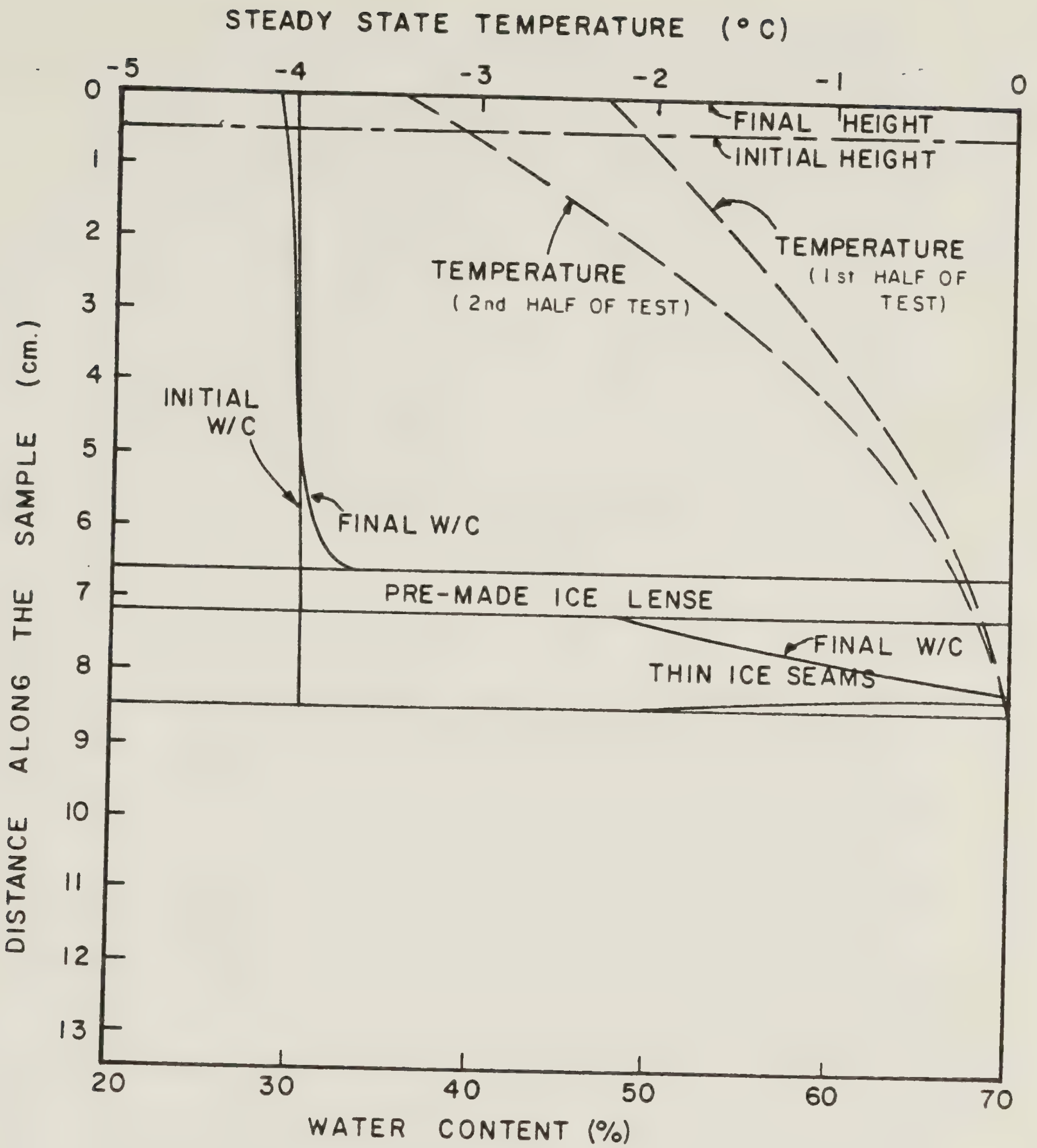
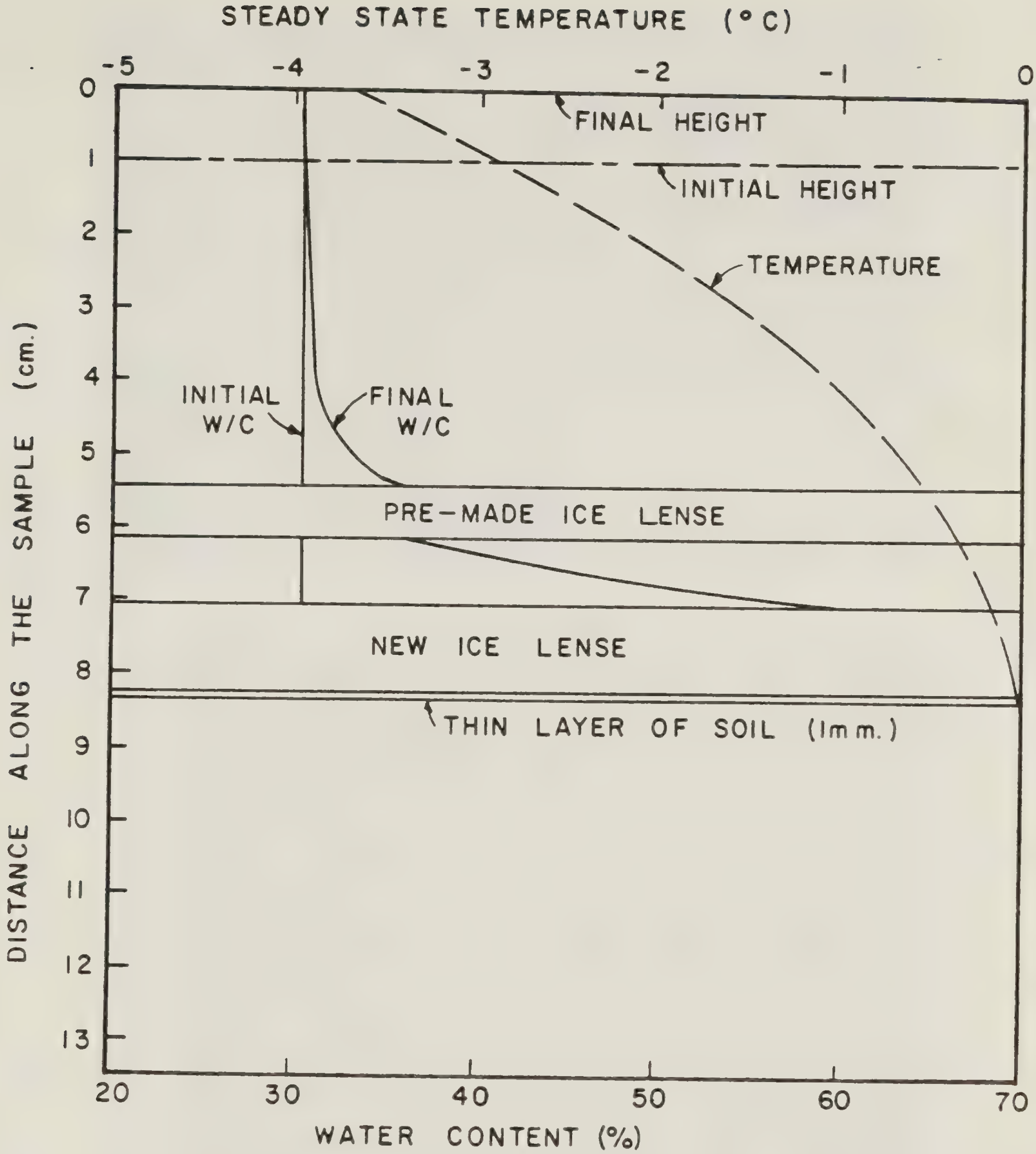



FIGURE 4.8 MOISTURE MIGRATION GRAPH FOR TEST A-9




| |
|---|
| OPEN SYSTEM |
| Δ TIME = 5.7 DAYS |
| Δ PRESSURE = 0.0 KPa. |

FIGURE 4.9 MOISTURE MIGRATION GRAPH FOR TEST A-12




OPEN SYSTEM

Δ TIME = 22 DAYS

Δ PRESSURE = 0.0 KPa.

FIGURE 4.10: MOISTURE MIGRATION GRAPH FOR TEST A-13


(A)

(B)

FIGURE 4.11 FROZEN DEVON SILT SAMPLE FROM TEST A-13 AFTER 22 DAYS.

(A) CROSS — SECTION OF SAMPLE. THE TOP ICE LENSE WAS PRE-MADE; THE BOTTOM LENSE FORMED DURING THE TEST.

(B) CLOSE-UP OF THE ICE LENSES. NOTE THE DECAY NEAR THE TOP OF THE PRE-MADE LENSE.

TABLE 4.1

Summary of Test Results

(Open System - No Applied Loads)

| Test No. | Length of Test (days) | Water Intake Rate (ml/hr) | Total Heave Rate (mm/hr) | Heave Characteristics |
|---|---|---|---|---|
| A-5 | 4.3 | 0.28 | 0.060 | -10mm Ice Lense with thin soil layer below. |
| A-6 | 5.0 | 0.11 | 0.005 | -One half of sample thawed and refroze. Reticulate ice only |
| A-7 | 5.8 | 0.19 | 0.028 | -5mm Ice Lense with no soil below. Ice formed on porous stone. |
| A-8 | 11.0 | 0.49 | 0.071 | -Pre-made Ice lense over 5mm soil. 15mm Ice Lense formed in this soil. |
| A-9A | 3.0 | 0.00 | 0.000 | -Pre-made Ice Lense with no soil below. No heave occurred. |
| A-9B | 11.0 | 0.00 | 0.000 | -A 1mm layer of soil was added to bottom of ice. No heave. |
| A-12 | 5.7 | 0.48 | 0.076 | -Pre-made Ice Lense with 9mm soil below Thin ice seams formed in this soil |
| A-13 | 22.0 | 0.54 | 0.078 | -Pre-made Ice Lense with 7mm soil below A 10mm ice lense formed in this soil |

## 4.4 Open System Tests - With Applied Loads

The testing program concluded with two open system tests involving the application of various loads. The purpose of these tests was to study the effects of overburden pressure on the location and amount of ice accumulation in the frozen zone.

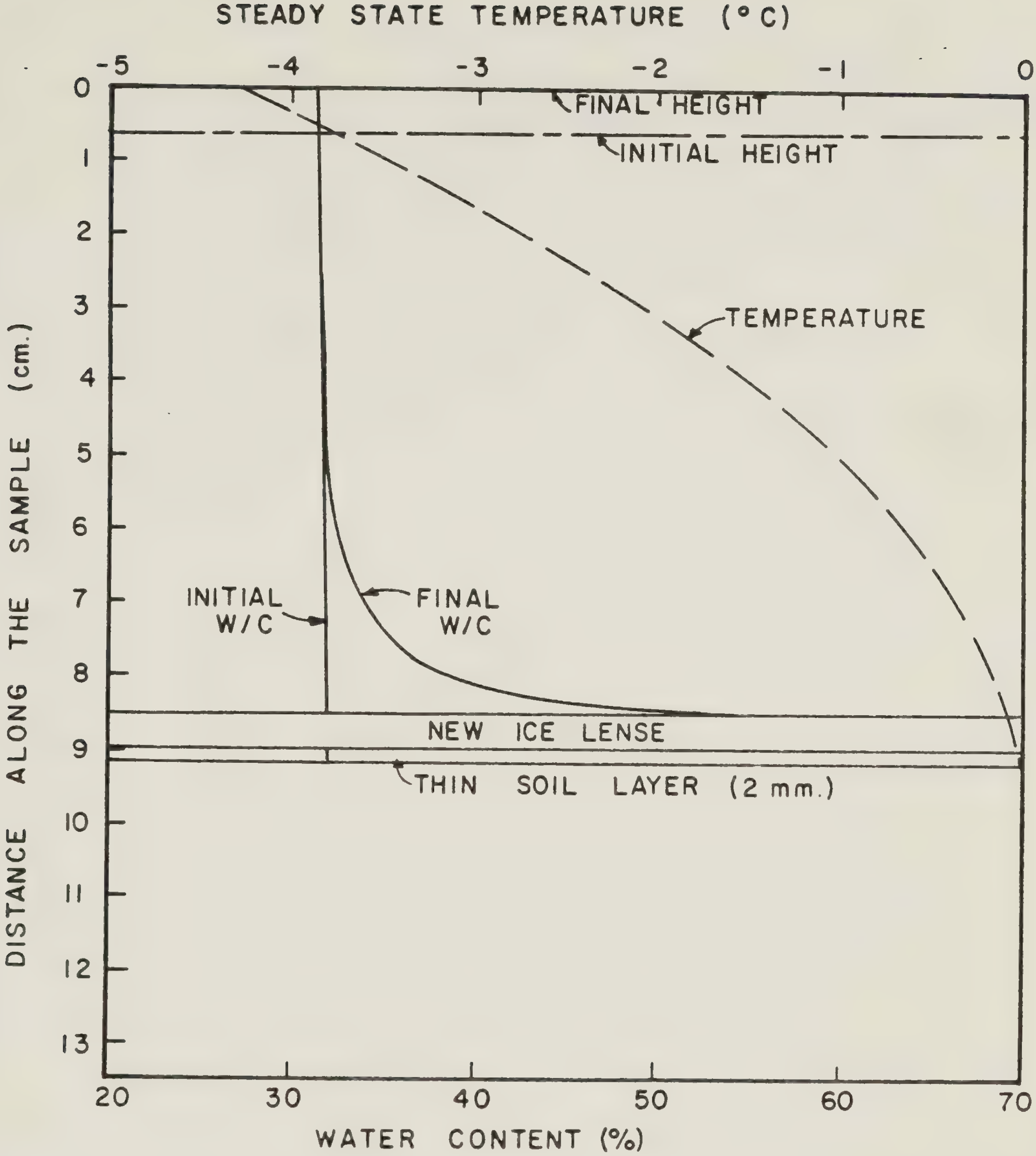
Tests A-10 and A-11 were conducted applied pressures of 50 kPa (time=5 days) and 75 kPa (time=6.5 days), respectively. Results of water content determinations are presented in Figures 4.12 and 4.13. Heave characteristics are shown graphically in Figures B-10 and B-11 and summarized in Table 4.2.

An examination of the results shows that an increase in water content within the frozen soil occurred in both samples. An ice lense developed in Test A-10 whereas only thin bands of ice were observed within the sample in Test A-11.

The sequence of loading was slightly different for the two tests. A pressure of 100 kPa was initially applied to the sample in Test A-10. After about 20 hours water intake still had not commenced. It was decided at this time that the load was too heavy to permit proper conditions for suction and frost heave to initially develop in the soil. Therefore, 50 kPa was removed and water intake immediately began. In Test A-11, pressure of 75 kPa was applied at the

start of the test. An initial expulsion of water due to consolidation in the warmer zones of the sample was noticed but after 10 hours the process was reversed and water began to be drawn into the sample.

An analysis of the location and rate of ice accumulation for these tests in comparison with that predicted by Penner and Walton (1978) as well as with results from Tests A-10 and A-11 is given in Chapter V.

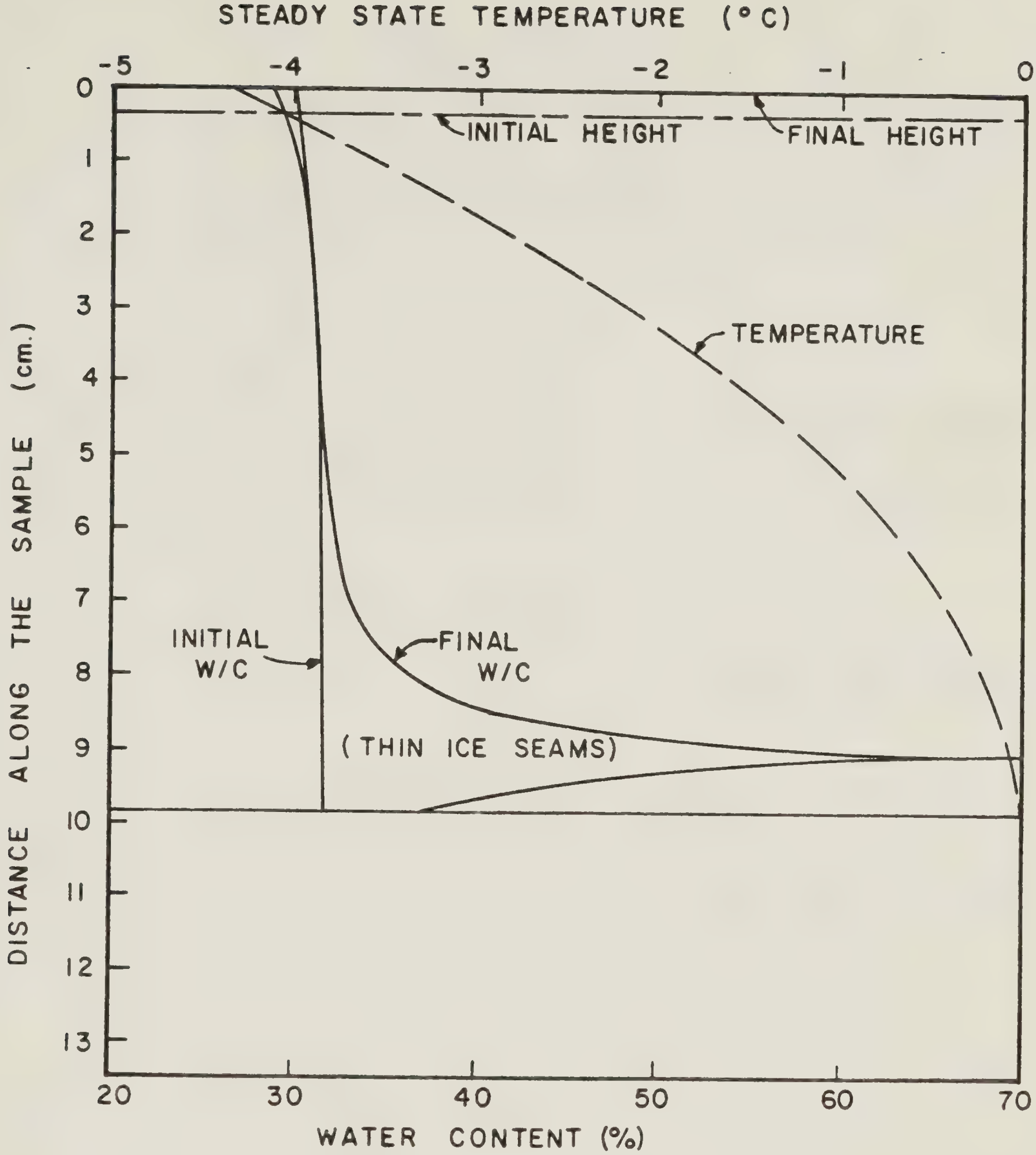



OPEN SYSTEM

Δ TIME = 5.0 DAYS

Δ PRESSURE = 50.0 K Pa.

FIGURE 4.12: MOISTURE MIGRATION GRAPH FOR TEST A-10




| |
|---|
| OPEN SYSTEM |
| Δ TIME = 6.5 DAYS |
| Δ PRESSURE = 75.0 KPa. |

FIGURE 4.13: MOISTURE MIGRATION GRAPH FOR TEST A-11

## TABLE 4.2

### Summary of Test Results

(Open System - With Applied Loads)

| Test No. | Length of Test (days) | Water Intake Rate (ml/hr) | Total Heave Rate (mm/hr) | Heave Characteristics |
|---|---|---|---|---|
| | Overburden Pressure = 50 kPa | | | |
| A-10 | 5.0 | 0.37 | 0.054 | -A 5mm Ice Lense formed with 2mm of soil below. |
| | Overburden Pressure = 100 kPa | | | |
| A-11 | 6.5 | 0.11 | 0.012 | -No ice lense formed Thin bands of ice were observed near the bottom. |

## 4.5 Summary of Test Results

A few general points can be made which are common to all test results:

1.) In all tests some migration of water occurred to varying degrees. In the closed system tests, a decrease in water content near the bottom of the sample and subsequent increase in adjacent zones was observed. The results from the open system tests, although not completely consistent with respect to magnitude, showed an increase in water content in the bottom zone of the frozen sample. The results from tests involving a pre-made ice lense revealed that some increase of water content in the soil above this lense. It is expected this transfer of moisture was the result of regelation.

2.) The extent of migration in the frozen soil corresponds closely to the point in the sample where the steady state temperature was near -2.0°C. Of course, the extent of migration also depended upon the length of testing, pressure effects and other factors which will be discussed more thoroughly in Chapter V.

3.) The rate of heave in nearly all the open system tests showed a gradual but significant decrease with time.

Explanations for this behaviour will also be presented in the following chapter.

# CHAPTER V

# Discussion

## 5.1 Moisture Migration in Frozen Soil

### 5.1.1 General Discussion of Test Results

The open system test results presented in Figures 4.4 through 4.12 provide valuable insight into the process of moisture migration in frozen soil.

It was not clear from the results of the first two open system tests (A-5 and A-7) whether incoming water migrated into the frozen soil prior to ice lense formation or during formation. However, the stages of moisture migration became clearer with the conclusion of open system tests involving pre-made ice lenses. It appears that water is initially drawn into the frozen soil prior to development of an ice lense. The incoming water moves up into the frozen zone at fairly high velocities due to the high suctions (see Figures 2.1 and 2.2). The permeability of the frozen soil at this stage is high enough to allow relatively free flow within the unfrozen water film. Water migrates upward and freezes within the soil pores. As more ice accumulates in the frozen soil an increase in tortuosity causes a net decrease in apparent permeability. Eventually it becomes easier for the water to freeze into segregational ice than to move up through the soil. Thus, the formation of an ice lense begins.

This scheme was deduced from the fact that in those tests involving a pre-made ice lense relatively little redistribution of moisture occurred behind this ice lense. In tests without a premade ice lense, however, the increase in water content within the frozen zone was significant. It appears that the process of regelation is less dominant in the earlier stages of frost heave. If regelation were of primary importance it would be expected that the redistribution of moisture behind the pre-made ice lense would be similar in magnitude to the redistribution behind the new ice lense in tests A-5 and A-7 since the temperature gradients were the same. This was not the case. It appears, then, that most of the moisture migration occurred within the unfrozen film of the soil and not by the upward movement of pore ice as suggested by Miller (1972). Moreover, the presence of an ice lense appears to impede this process.

### 5.1.2 Results from Tritium Analysis

The purpose of utilizing the technique of isotope tracing in the present study was to attempt to identify where and how water migration occurs in frozen soil. The velocity of flow in frozen soil was also a factor which was of interest.

Typical results from tritium analyses along with initial and final water content profiles for three tests are

presented in Figure 5.1.

Figure 5.1 (A) represents the results from Test A-6 where approximately one half the sample had thawed and subsequently refroze. The amount of new (tritiated) water in the soil decreased rapidly from almost 100% near the bottom to only about 8% at a distance of 3.2 cm from the bottom.

The general shape and magnitude of the tritium curve for Test A-10 (Figure 5.1B) is very similar to that of Test A-6. In Test A-10, a load of 50 kPa was applied to the frozen soil and water was allowed to be drawn in. An ice lense about 0.5 cm thick had developed within the frozen soil prior to testing, are presented in Figure 5.1(C). It is evident that very little tritium passed through the pre-made lense. In fact, the quantity measured in the soil immediately above the lense could probably be attributed to contamination. However, a significant tritium count was recorded for the soil layer in between the two ice lenses. This corresponds to the increase in water content of this soil layer. The new ice lense would be expected to be composed entirely of tritiated water, yet the results show a tritium content of less than 100%. This can be explained by the fact the water from the thin layer of soil below the new ice lense was included in the analysis causing some dilution of the tritium content.

In conclusion, these tests provide empirical evidence

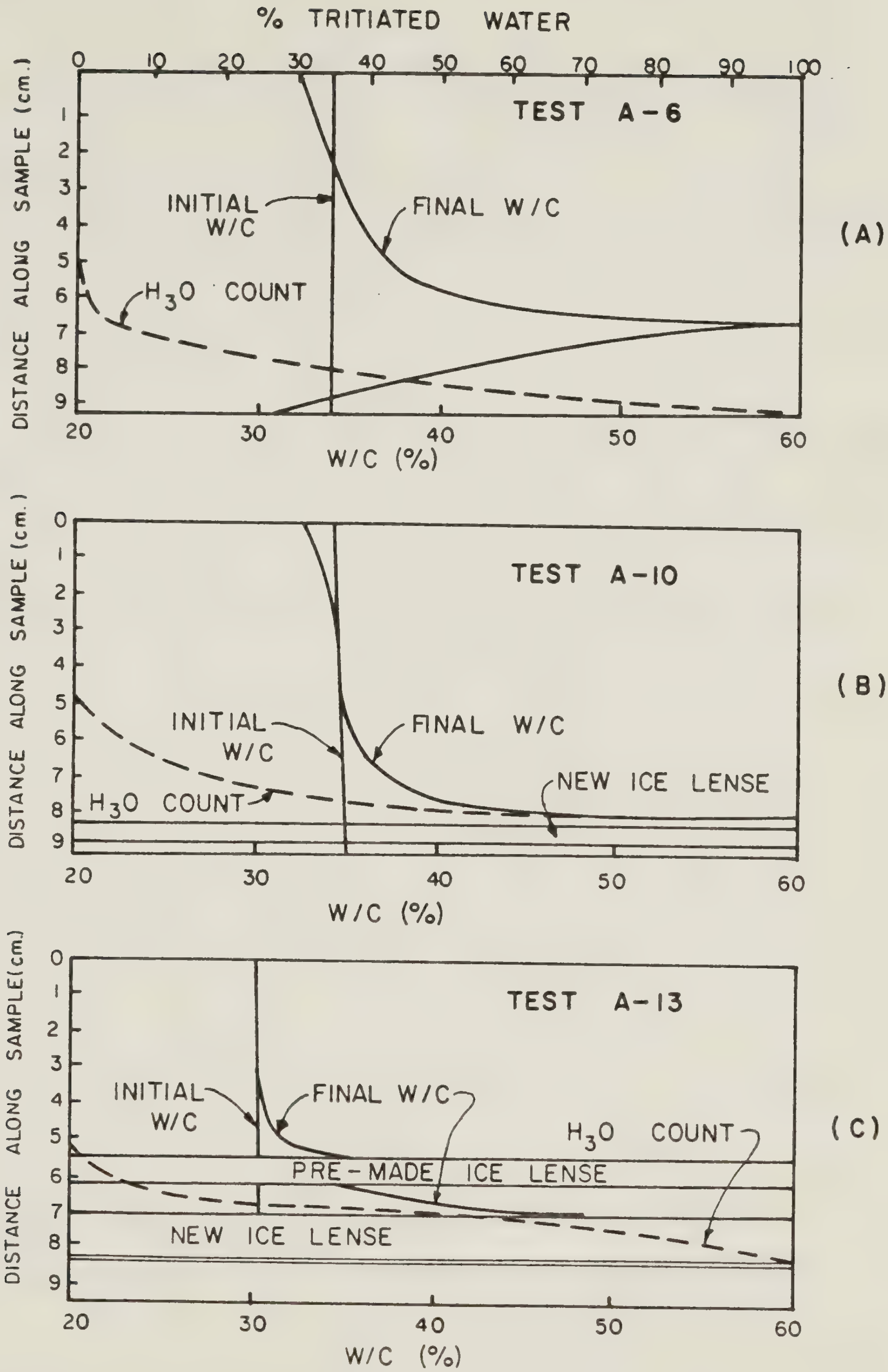



FIGURE 5.1 RESULTS FROM TRITIUM ANALYSIS FOR A FREEZING SOIL (A), FROZEN SOIL WITHOUT A PRE-MADE ICE LENSE (B), FROZEN SOIL WITH A PRE-MADE ICE LENSE (C).

that:

i) Moisture moves more freely through frozen soil without an ice lense.

ii) Once an ice lense forms the new water collects at the bottom of the lense with little transfer through the lense.

An estimate of migration velocities can be deduced if the boundary of the tritiated water can be properly identified. A procedure was developed to ascertain the limit of migration of tritiated water in the thawed portion of Test A-6. The results shown in Figure 5.1 were replotted against a normalized depth in Figure 5.2(A). From this plot an H3O distribution curve can be developed in which the percent of tritiated water is related to the normalized depth. This is shown in Figure 5.2(B). A regression analysis resulted in a mean average distribution occurring at a normalized depth of 65%. This corresponds to a distance of 1.3 cm from the bottom of the sample. If it is assumed that most of the new water migrated into the soil during the thawing and refreezing process, a time for this amount of migration to occur can be obtained from Figure B-6. A time of 25 hours is chosen from the volume change vs. time plot of Figure B-6. The corresponding rate of water migration would be 1.3 cm/25 hr = 0.052 cm/hr. However, the average rate of heave measured during this period (as determined by the change in water intake) was only 0.015cm/hr. This value

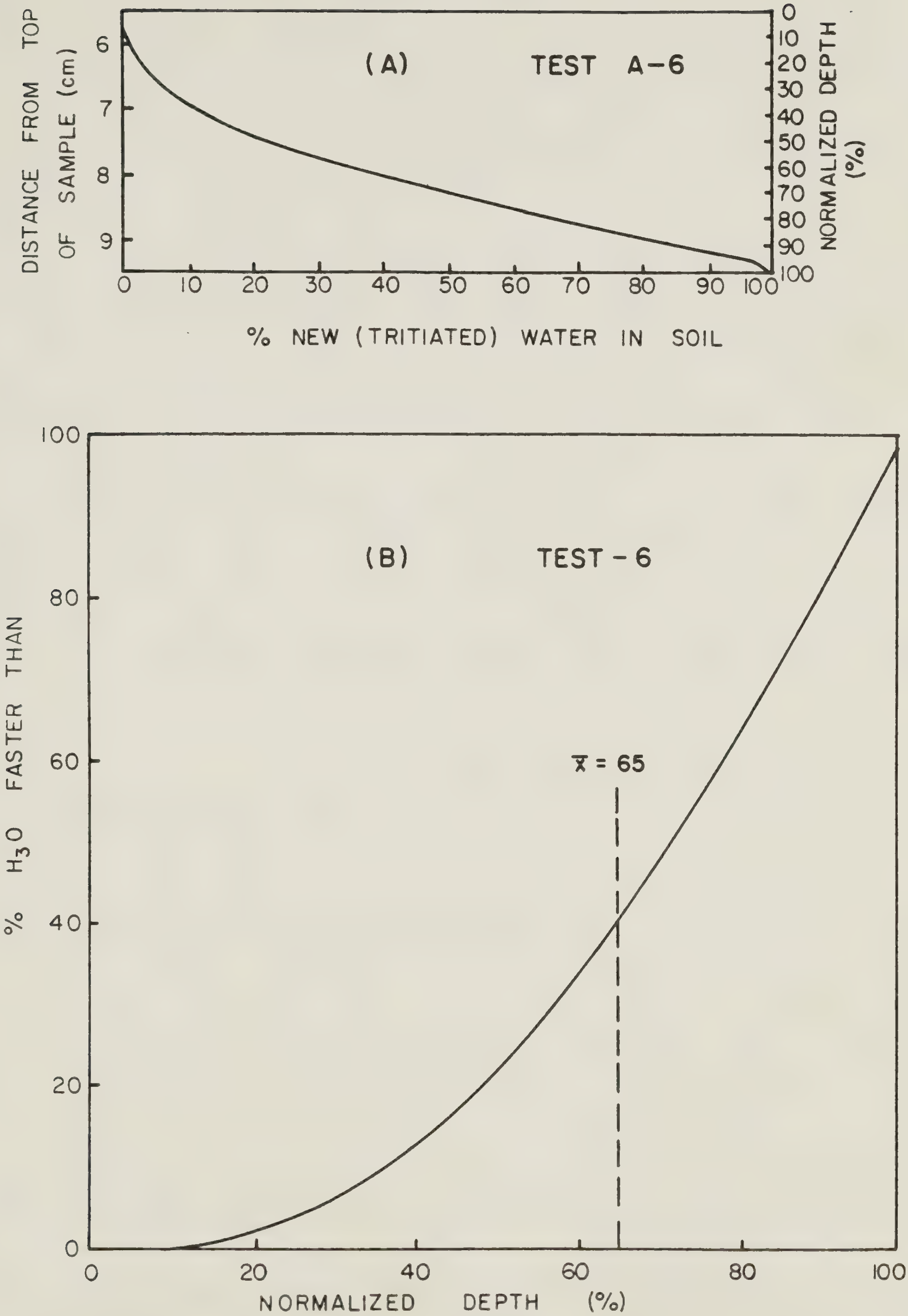



FIGURE 5.2 RELATION OF THE AMOUNT OF TRITIATED WATER TO NORMALIZED DEPTH FOR THE PURPOSE OF CALCULATING MIGRATION VELOCITY.

would result from the tritium analysis only if the extent of tritiated water migration was taken to be 0.4 cm. It is suggested, therefore, that the migration of tritiated water in the refrozen zone occurred throughout the duration of the experiment. Assuming this to be true the resulting average velocity of moisture migration (during the time the soil was thawed and during the time it was frozen) would be 1.3 cm/125 hr = 0.010 cm/hr. This value is closer to the measured rate of water migration.

The above analysis was designed more as an illustrative exercise than as a method for determining the actual velocity of moisture in the soil although with good quality data this would be possible. For the tests presented in this thesis, however, there are too many factors influencing the results for a precise quantitative analysis to be undertaken. This is especially true for those tests where tritium was used in a completely frozen soil. Many difficulties including contamination led to results which were less than adequate for an accurate evaluation of the extent of tritium migration. Also, an estimate of a value for the time for a certain amount of migration to occur is complicated by the fact that the processes involved are not well understood. Thus, it is not known when and how much the moisture migration is impeded by the formation of an ice lense in the frozen zone. Future tests could be designed to examine this behavior.

### 5.1.3 Ice Accumulation in Frozen Soil

The literature review in Chapter II included a synopsis of recent work by Penner and Walton (1978). A more in depth treatment of their hypothesis regarding rates of ice accumulation in frozen soil will be given here.

Penner and Walton empirically obtained a relation between heave rate and pressure and temperature (Equation 2.8) from open system frost heave tests. By differentiating Equation 2.8 with respect to temperature they obtained a relation between the change in rate of ice accumulation (heave rate) per degree Celsius and pressure and temperature. This is given in a generalized form in Equation 5.1.

$$\frac{d(dh/dt)}{dT} = c \cdot (P/T^2) \cdot e^{d(P/T)} \qquad (5.1)$$

where: $dh/dt$ = heave rate
$P$ = overburden pressure
$T$ = temperature in the frozen soil
$c$ and $d$ are constants and depend on soil type

A plot of Equation 5.1 was given in Figure 2.5 for a clayey silt material. The authors had no physical evidence for the predictive ability of Equation 5.1. The results presented in Chapter IV provide an opportunity to test Penner and Walton's prediction. The results from three tests (A-5, A-10 and A-11) were analyzed. For each test the change in water content with time was calculated for incremental slices along the sample. This value was then divided by the

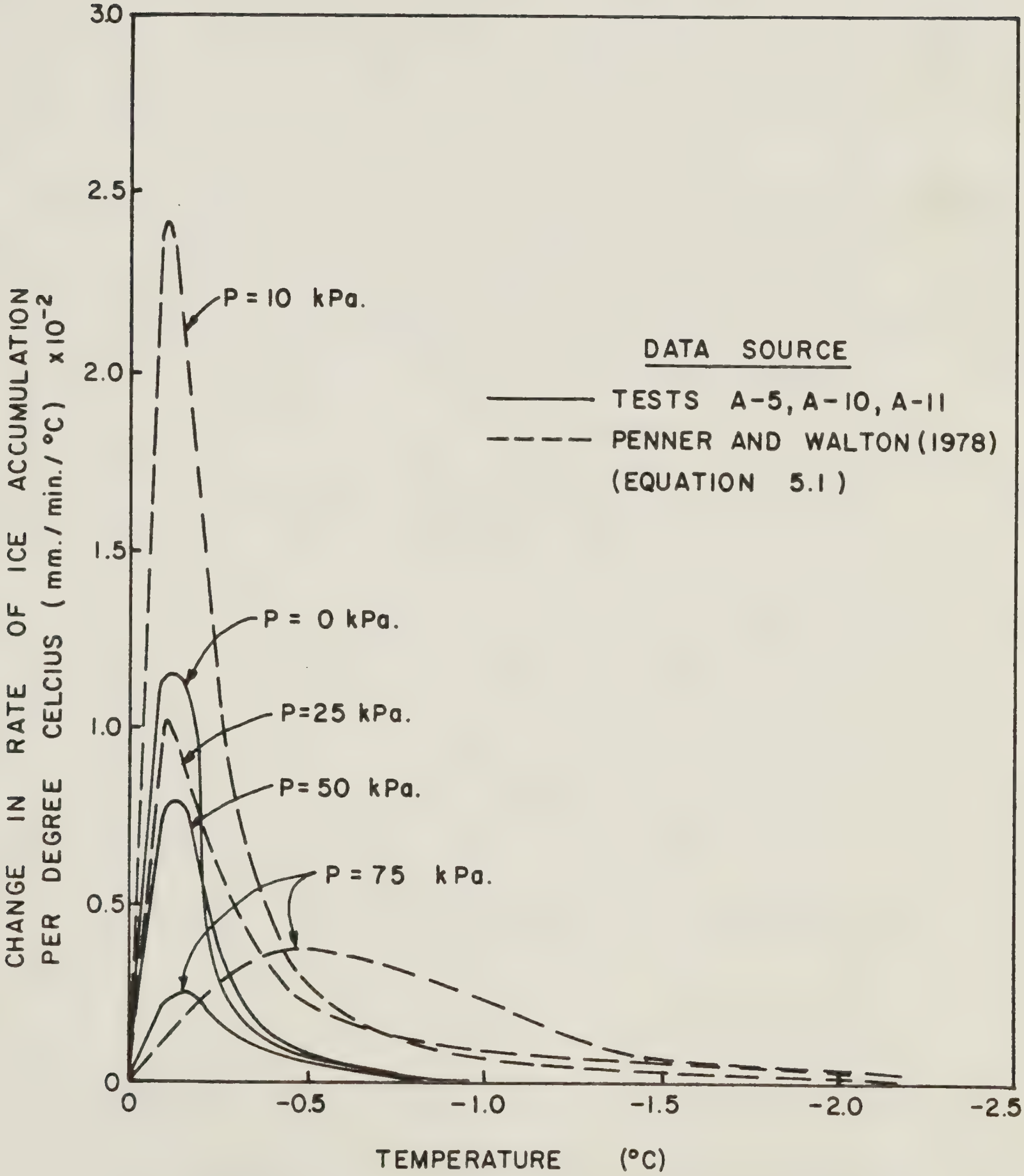



FIGURE 5.3 COMPARISON OF RATE OF CHANGE OF ICE ACCUMULATION IN FROZEN SOIL.

average temperature of the corresponding slice. A plot of the results along with Penner and Walton's values for clayey silt based on Equation 5.1 is given in Figure 5.3. The values of $c = -4.26 \times 10^{-3}$ and $d = 0.968$ were used in the construction of the plot and were obtained from Penner and Walton (1978).

Although magnitudes are slightly different the general shapes of the curves are consistent. As predicted by Penner and Walton, more ice accumulates near 0°C at low overburden pressures. At higher pressures ice accumulation moves further into the colder zones. Therefore, besides reducing heave rates, overburden pressure causes the zone of heaving to increase over greater distance. This is to be expected since it was seen in Chapter II that one effect of pressure is to increase the unfrozen film thickness. An increased film thickness in the colder regions reduces the potential to draw in water but it also increases the mobility of water molecules within the film. Thus, less water would be drawn into the frozen soil but the extent of migration would be greater.

### 5.1.4 Permeability of Frozen Soil

Frozen permeability calculations were performed for Test A-5. The analysis involved plotting a suction profile as a function of temperature throughout the frozen zone as per the theoretical relationship given by Harlan (1974) (see

Figure 2.1). This is shown in Figure 5.4. The figure was divided into 9 sections of one centimetre thickness (plus an additional 0.5 cm section). The average temperature of each section and the change in suction from the top and bottom of the section was obtained from Figure 5.4. It was assumed that the rate of water inflow as shown in Figure B-5 represented the rate of moisture migration into the frozen zone. If the conservation of mass is to be satisfied, this assumption must be true. Therefore, since the suction gradient and the velocity of water flow for each section is known, a calculation of the corresponding permeability based on Darcy's Law can be made. The results are shown graphically in Figure 5.5. Also included in Figure 5.5 is a plot of the results from tests by Williams and Burt (1974) and Miller (1970 ). Although the values for permeability given by Williams and Burt are generally higher than those reported by Miller or calculated from the results of Test A-5, a reasonable correlation is seen. It is especially interesting to note the shape of the plot for permeability of frozen Devon silt (Test A-5). A distinct change in permeability decreases at a much slower rate with decreasing temperature. Also the slope of the plot flattens out near the temperature, T = -1.5°C. It was mentioned in Chapter IV that the limit of observable change of water content appeared to correspond to this temperature in the soil. This behavior probably results from the low apparent permeability of frozen silt at temperatures below -1.5°C. This phenomenon

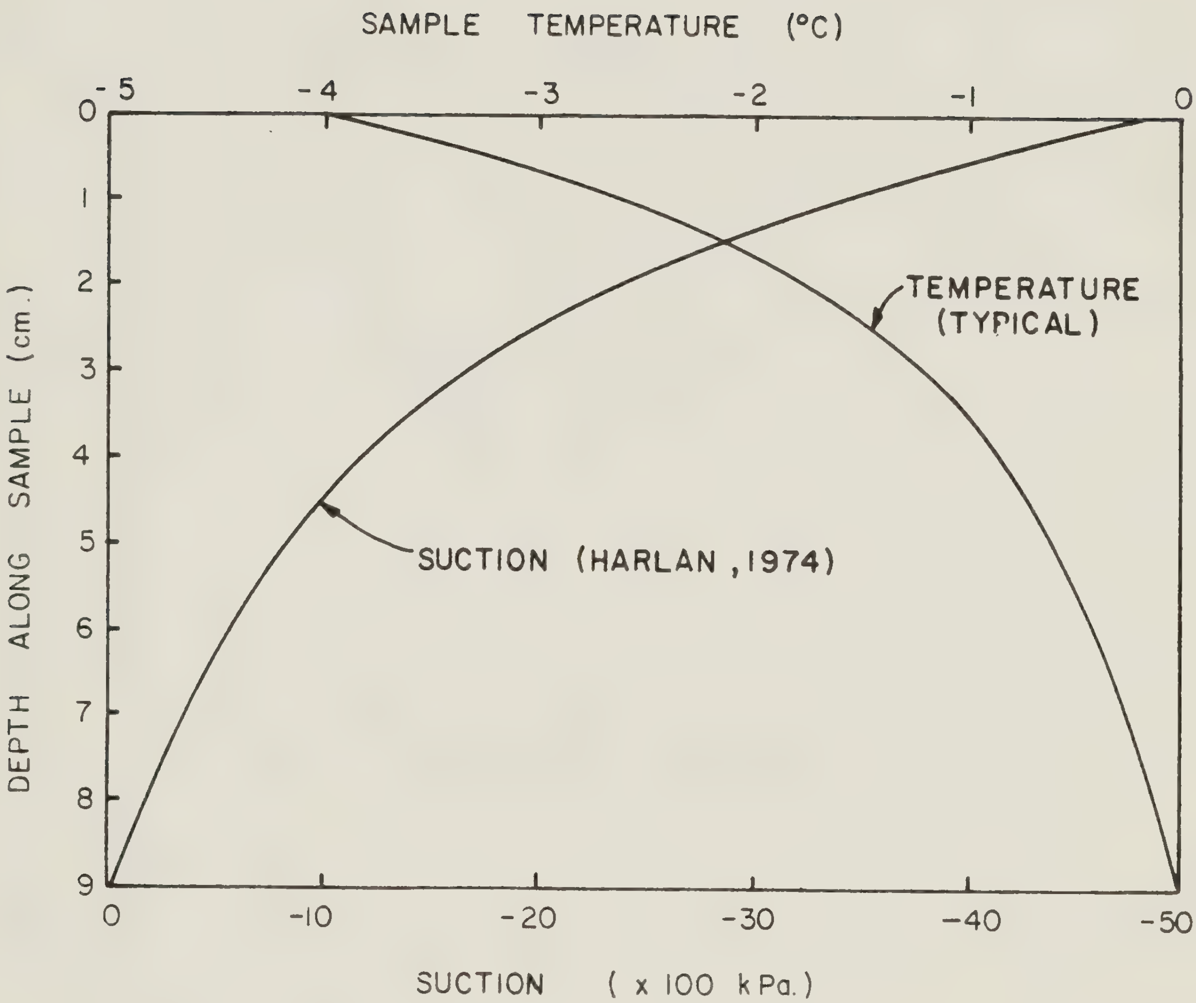



FIGURE 5.4 SUCTION PROFILE BASED ON A THEORETICAL RELATIONSHIP. (HARLAN, 1974)

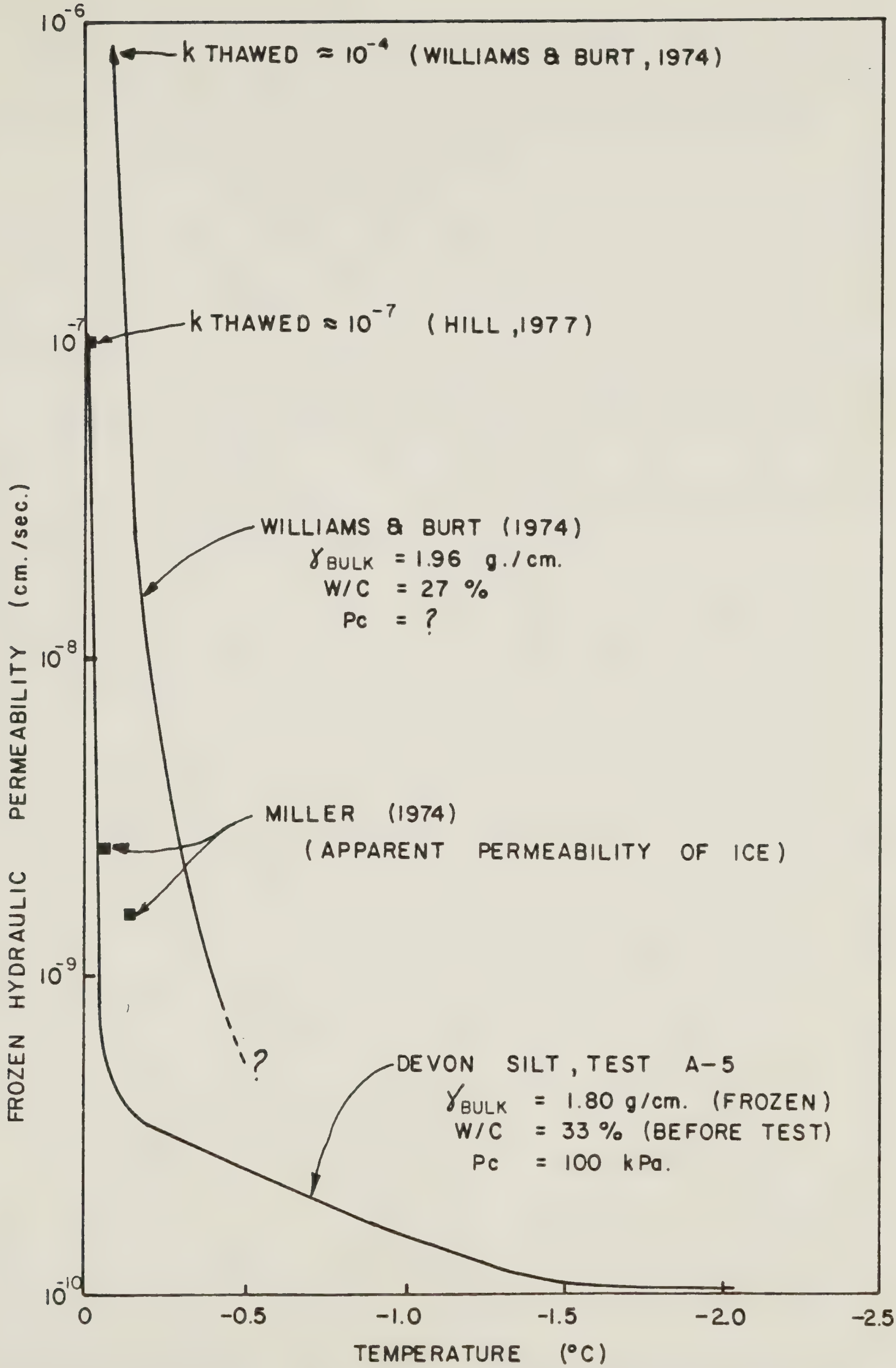



FIGURE 5.5 FROZEN PERMEABILITY vs. TEMPERATURE

could be better examined with the development of an unfrozen water content-temperature curve for the Devon silt.

It should be noted here that a change in heave and water intake rate was observed in Test A-5 (Figure B-5). Since it has been shown in Section 5.1.1 that the increase in water content of the frozen soil probably took place prior to ice lense initiation, it can be concluded that the suction pressures are initiated within the frozen fringe and are not influenced to any great degree by the amount or location of ice accumulation. Moreover, the governing factor in the control of water intake rates (and, therefore, heave rates) are most likely determined within the frozen fringe of the soil below the lowest (warmest) ice lense.

The effect of the suction force within this fringe on water intake is apparently dependent upon the configuration of the ice-particle interphase below the ice lense. In Tests A-9A and A-9B no heave was observed although the temperature and boundary conditions were very similar to those of Test A-7 where a significant ice lense formed. It can be seen from Table 4.1 that water intake and heave rates are very inconsistent. This is attributed to the sensitivity of heave rate to the boundary conditions below the warmest ice lense. It is possible, of course, that the frost cell could be of some significance for tests with low overburden pressure, thus contributing to the variability of test results. In nearly all the open system tests an apparent decrease of

heave rate with time was observed. The physics behind this phenomenon are unclear but a general interpretation of the processes involved is presented in Section 5.2.2.

### 5.1.5 Summary of Discussion on Test Results

From the test results presented in Chapter IV and the discussion above several important observations can be made pertaining to migration of water in frozen soil.

1) Water does migrate through frozen soil under the influence of a temperature gradient.

2) The magnitude and extent of migration appears to be closely controlled by the apparent permeability of the frozen soil. The frozen permeability for a given soil is a function of several factors including temperature (which determines the amount of unfrozen water), amount of ice accumulation, and the overburden pressure. It is evident that for at least a certain time period, the migration of moisture in a frozen soil is impeded by the accumulation of ice structures in the frozen zone.

3) The significance of a regelation process has not been adequately asessed; however, it appears that compared to water transport through the unfrozen film its

contribution would be small, at least for short time periods. The time effects of frost heave will be discussed further in Section 5.2.2.

4) One effect of pressure on a frozen soil is to create a greater percentage of unfrozen water in the frozen soil at a given temperature. This is illustrated by test results from experiments with applied overburden pressure which showed that ice accumulates further into the frozen zone with increased pressure.

5) Although a suction does exist throughout the frozen zone it appears that the seat of suction potential for causing frost heave is located within a small zone adjacent to the frozen-unfrozen interphase. The effect of the suction deeper within the frozen zone is expected to cause some redistribution. However, it is thought that the influence of this redistribution to the rate of heave would be small within an engineering time frame. This is due to the fact that the mechanism for moisture movement after an ice lense has formed is mainly regelation. It has been shown that this process is very slow. The movement of moisture through the unfrozen water film would be extremely limited since the permeability in the coder zones is very low.

## 5.2 General Discussion of Frost Heave

### 5.2.1 Steady State Frost Heave

In the open system tests discussed in Chapter IV the experimental setup was designed in an attempt to simulate field conditions where the frost front has stopped advancing. This condition is termed "steady state" frost heave. It is of interest to examine the possibility of steady state frost heave occurring in the field situation within an engineering frame of time.

It is well known that the depth of frost penetration can be predicted from the semi-empirical relation:

$$X = \alpha / \sqrt{t} \qquad (5.2)$$

where: $X$ = frost penetration (m)
$\alpha$ = freezing parameter ($m/hr^{1/2}$)
$t$ = time (hr)

Equation 5.2 has been tested over a two year period in a field frost heave test site (Slusarchuk, et al, 1973). The predicted penetration rate correlated well with actual field measurements. However, it cannot be expected that the penetration of frost would continue indefinately as implied by Equation 5.2. The upward flow of moisture, a continual increase in overburden pressure and a decrease in temperature gradient would eventually cause the frost front to stop advancing. A typical field situation is shown schematically in Figure 5.6.

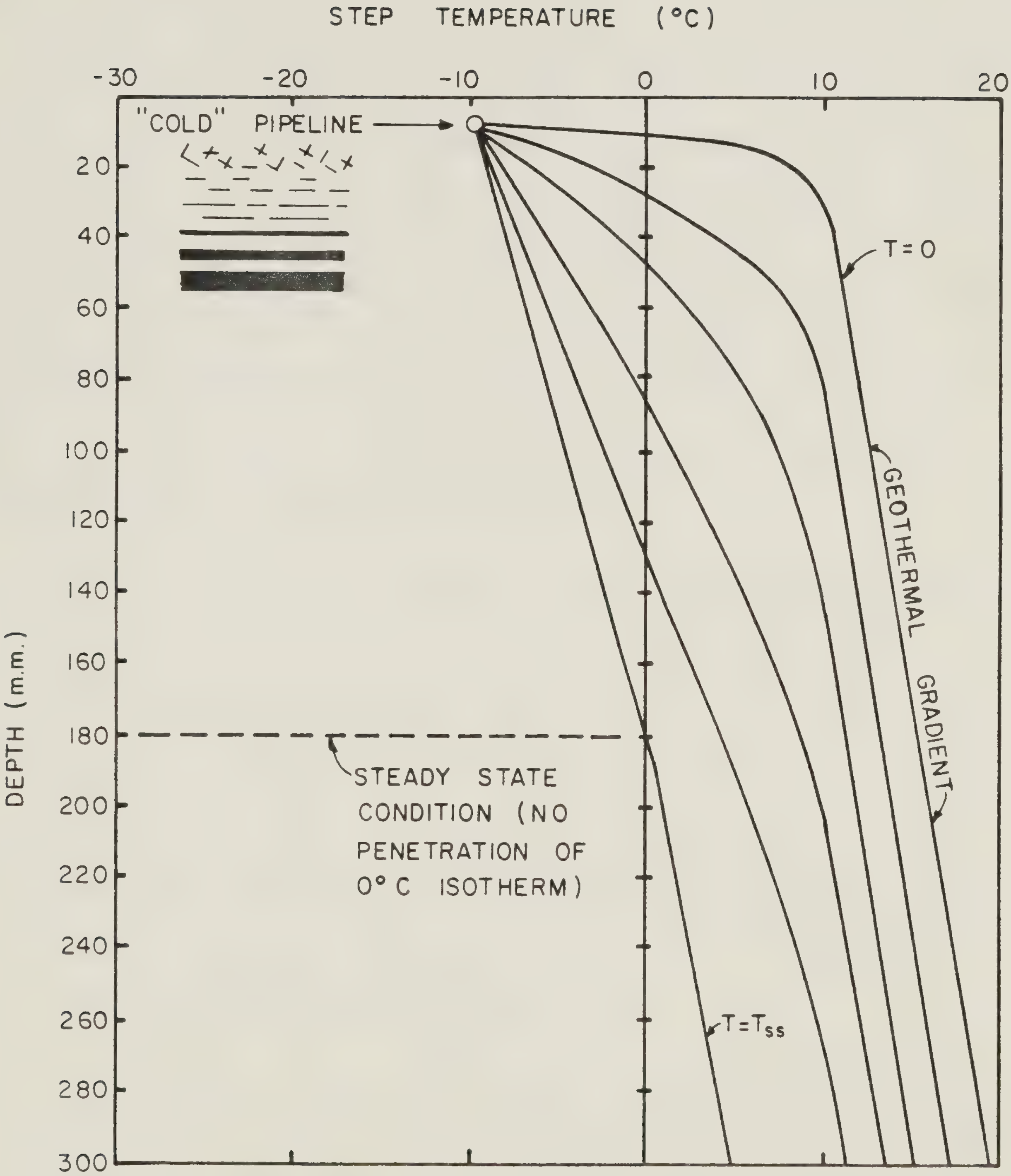



FIGURE 5.6: SCHEMATIC DIAGRAM OF FROST PENETRATION WITH TIME IN A TYPICAL FIELD SITUATION.

Assuming an initial ground temperature of -10°C at the surface and a geothermal gradient of 1°C/30m, it is seen that steady state conditions with respect to frost penetration would develop at depth on the order of 180 meters. This is neglecting the contribution of sensible heat released from water drawn to the frost front and the important effects of overburden pressure. Advanced computer techniques could be developed to evaluate the time at which this condition would actually occur but it is obvious without a complicated analysis that the time to steady state conditions is beyond the realm of ordinary engineering design periods (typically 30 years).

Although true steady state conditions are never achieved within typical design periods, quasi-steady state conditions are present at various stages of frost heave. In Figure 5.6, characteristic ice formations are shown at various depths below the bottom of a cold pipe (T = -10°). A qualitative description of the processes involved with these types of formations will provide insight into the meaning of a quasi-steady state condition.

Initially, the frost front moves rapidly through the soil due to the high temperature gradient. Water is drawn to (and behind) the frost front as result of the temperature-induced potential gradient within the warmer part of the frozen zone. The rate of flow of water is limited by the magnitude of suction and permeability of the unfrozen soil

according to Darcy's Law. The removal of heat near the surface is so rapid and the water is frozen so quickly that few ice lenses are formed. A reticulate ice structure first develops and gradually changes to horizontal bands of thin ice seams increasing in thickness with this depth. This phenomenon was observed in Test A-6. Hill (1977) found a similar ice structure in several of his tests where a cold step temperature was used. This process represents the condition of a frost front that is constantly advancing with virtually no delays for ice development.

As the frost front penetrates further the temperature gradient in the frozen zone gradually decreases. This produces two interconnected events. Firstly, the rate of penetration decreases since there is less removal of heat at lower depths. Secondly, as a result of slower frost penetration rates, the water drawn to the frost front has time to develop into thicker ice seams. A considerable amount of latent heat is released when water is converted into ice (79 cal/g). As the rate of heat removal is lowered (with increasing depth) the latent heat has a continually greater effect on the rate of frost penetration. When an ice lense begins to form due to some localized disturbance in the heat-mass balance, part of the heat removal which was previously causing the frost front to advance is used to remove the latent heat of fusion of water. Hence, a temporary interruption of the frost penetration results.

This temporary state, then, is termed quasi-steady state frost heave.

The fact that the rate of frost penetration is related to the rate of ice formation has been shown in the field by Slusarchuk, et al (1978). The total amount of heave at a certain time (2.5 years) for an unrestrained cold pipeline was approximately one-half of that for a restrained pipe. However, the depth of frost penetration was 20% less for the unrestrained pipe. Laboratory experiments by Penner and Ueda (1977) showed similar behavior. This evidence clearly demonstrates that when the heat removal rate is constant the frost front will penetrate slower during ice lense formation.

The rate of water intake during this quasi-steady state condition is a function of the rate at which the latent heat of fusion can be transported up through the frozen soil. Assuming the sensible heat from the new water migrating to the frost front is negligible, the conservation of energy at the 0°C isotherm is given by:

$$Q\ in = Q\ out \qquad (5.3)$$

where:

$$Qin = L(h/t)\cdot A + Ku(\partial T/\partial x)u\cdot A \qquad (5.4)$$

$$Qout = kf(\partial T/\partial x)\cdot A \qquad (5.5)$$

and: $Q$ = heat flow
$L$ = latent heat
$h/t$ = heave rate
$(\partial T/\partial x)$ = temperature gradient
$k$ = thermoconductivity
$A$ = area
u, f refer to the frozen and unfrozen states

By equating Equations 5.4 and 5.5 and including a factor for the volumetric expansion due to the freezing of water, an expression for heave rate during ice lense formation is obtained:

$$h/t = \frac{kf(\partial T/\partial x)f - ku(\partial T/\partial x)u}{L} \qquad (5.6)$$

Equation 5.6 represents the upper bound of heave rate since it was tacitly assumed that the permeability of the unfrozen soil would not impede ice lense growth. Also, no provision is made to include the effects of pressure on the system.

It is important to note that $(\partial T/\partial x)f$ in Equation 5.6 refers to that gradient within the frozen fringe. For a field situation the gradient within the fringe should be very close to the overall temperature gradient in the frozen soil except for slight variations due to localized microscopic differences in thermoconductivity. In laboratory experiments, however, the gradient in the frozen zone was not linear due to the effects from the sides of the apparatus. This caused the temperature gradient in the frozen fringe to be much higher than the overall gradient in the frozen zone. Therefore, a difficulty arises when an attempt is made to apply Equation 5.6 to the test results in

that the actual temperature gradient in the fringe must be interpolated from insufficient temperature data.

The second term of Equation 5.4 represents the heat in the unfrozen ground due to the geothermal gradient. For laboratory conditions this term can be replaced by the value of the heat flux into the soil from the warm (bottom) plate. Unfortunately, the heat flux from the plate was not measured, introducing additional unknowns into Equation 5.6.

The theoretical conditions for quasi-steady state frost heave in the field have been presented. How well the experimental tests represent this condition is difficult to ascertain. It is felt that the boundary conditions in the experimental setup may produce different behavior in the test sample than would be experienced in the field. The temperature gradient in the frozen zone (approximately 0.4°C/cm or 40°C/m) is thought to be much higher than would anticipated in normal field situations after the initial freezing period. It is important to consider these factors in the design of future tests.

### 5.2.2 Time Dependence of Heave Rate

Laboratory test results from Hill (1977), Penner and Ueda (1978) and from Chapter IV of this thesis indicate that heave rates tend to decrease with time. Field evidence of this phenomenon was provided by Slusarchuk, et al (1978). A

comprehensive explanation for this behavior is not yet at hand; however, it is of interest to discuss in more detail some relevant literature. A discussion of the important factors affecting heave rate will also be presented in this section.

In laboratory experiments dealing with the effects of pressure, temperature and time on heave rates, Penner and Ueda (1977) found very little decrease in heave rate with time even though the frost front had nearly penetrated the entire sample. It was suggested in Chapter II that the reason for such small time effects was that the test represented the situation of the early stages of frost heave where the frost front is advancing at a very high rate.

The temperature gradient in the frozen zone at the maximum penetration of the 0°C isotherm was about 0.18°C/cm or 18°C/m. For the field frost heave test reported by Slusarchuk, et al (1978) this gradient occurred after only a few months of testing which would still be considered within the early, more transient stages of frost heave.

Penner and Walton (1978) attempted to explain the decrease of heave rate with time observed in later long-term tests. They postulated that heave rate was a function of the ice segregation ratio or the total heave heave over the total frost penetration (Equation 2.9). In the development of this theory they assumed that heave involves the entire

frozen zone and that heaving would stop only after the total amount of heave equaled the total frost penetration. This perception is considered useful in that it brings to light the importance of time effects on heave rate. However, the relationship proposed is not based on any physical process occurring within the soil. Also, insufficient data does not allow the long-term predictive capability of their relation to be tested adequately. It is important, therefore, to develop a more complete understanding of the physical processes involved before the construction of a predictive model is attempted. A discussion of several factors believed to influence heave rate is given below.

It is suggested that, for a given fully saturated soil, there four factors which most significantly affect the rate of heave. They are:

i) $\partial T/\partial x$ = temperature gradient in the frozen fringe

ii) $P$ = overburden pressure

iii) $e$ = void ratio of the unfrozen soil (which determines the unfrozen permeability)

iv) $dw$ = depth to water table

The degree to which each of these parameters affects heave rate is not known at this time but a qualitative description of how each changes with time (or depth of frost penetration) and the subsequent effect on heave rate is given.

It is proposed that a certain suction develops within the frozen fringe of a freezing soil under given initial conditions. The magnitude of this initial suction is governed by the temperature gradient in the fringe, overburden pressure and void ratio. It is believed that a certain limiting suction at the interphase exists for a given soil and is a function of freezing point depression of the soil. However, for this discussion it is convenient to bypass the complexities of this point and to merely assume that a certain initial heave rate is caused by the development of an initial suction pressure at or within the ice-water interphase.

As the frost front penetrates several changes occur within the soil which would lead to a decrease in heave rate. Firstly, the temperature gradient within the frozen fringe decreases since the frost front is further from the cold source. Secondly, the overburden pressures increase due to an increase in the thickness of the frozen soil. Thirdly, assuming the water table did not fluctuate, a larger positive pressure would result at the frost front with increasing depth.

At the same time, the void ratio and, therefore, permeability in the unfrozen soil would be decreasing due to the increase in overburden pressure. The effect of smaller voids on heave rate is difficult to analyze. According to

the Kelvin equation (Equation 2.2) a higher suction should develop which would result in a higher heave rate. However, the lower permeability would inhibit the flow of water, thus reducing the heave rate. In addtion, excess pore pressure could develop from the consolidation process. This depends somewhat on the rate of advancement of the frost front.

It is evident that the influence of the various factors on frost heave rate is difficult to assess. Moreover, the interrelationship of the factors and their net effect on heave rate with time is even more complex. Therefore, an explanation for the observed decrease in heave rate with time, for both laboratory and field tests, is only speculative at this time.

It is expected that the decrease in heave rate with time reported by Slusarchuk, et al (1978) for an insitu frost heave experiment can be attributed to decreased temperature gradient in the frozen zone and to the increased overburden pressure with time. Which factor was most influential in reducing the heave rate is not known. A change in void ratio probably occurred but it is thought that the effect on heave rate was not as significant as the two above mentioned parameters. The water table fluctuated little (within 0.3 m) and appeared to have had little influence on heave rate.

It is suspected that the contribution of moisture

migration and redistribution through the frozen soil to the overall heave rate would be minimal for engineering design periods. The field tests reported by Slusarchuk, et al (1978) indicated that little or no observable heave had occurred within the frozen zone during a two year period. The effect of moisture migration in the frozen soil (secondary heave) may, however, be significant over long periods of time. The process of regelation behind the frost front is very slow and, although it is an ongoing process through all stages of frost heave, it would contribute little to the heave rate directly over a short period of time. Within a geologic time frame this process would cause a redistribution of ice throughout the frozen zone which in turn would change the heave characteristics of the soil. The net effect is not clear but an increased heave potential would probably result.

The phenomenon of heave rate reduction with time in laboratory tests is more difficult to explain. The results presented in Chapter IV indicate a definite reduction of heave rate with time in all open system tests although the temperature at the top and bottom of the sample remained constant throughout almost every test. The side temperature fluctuated for most tests but no correlation can be made with the observed decrease in heave rate. As well, it was proposed in Section 5.1 that the suction pressure is developed within the frozen fringe near the bottom of the

frozen zone; thus, the influence of side temperature on water intake from the bottom would be minimal.

A certain temperature equalization period is required for the frozen sample to achieve a steady state condition. The equalization period depends on the initial sample temperature, height of sample and the imposed steady state temperature gradient. Based on observed changes in temperature and theoretical calculations it is estimated that the maximum time for temperature equalization would be in the range of several hours. It is expected, then, that the contribution of temperature equalization on heave rate reduction may be significant during the first few hours of testing. The most significant reduction in heave rate, however, usually occurred many hours after this. Therefore, some other change in either the boundary conditions or in the sample iteslf (or both) must have occurred throughout the test.

Several interpretations are presented in an attempt to explain the observed behavior. During the removal of the frozen sample at the end of a test, it was noticed that, in most cases, some resistance was encountered. This resistance was attributed to the pressure of the ice lense against the sides of the frost cell. Side friction was not measured. However, the magnitude of such friction was probably great enough to restrict the upward movement of the sample to some degree as the ice lense developed.

The change in configuration of the frozen fringe could possibly have influenced the heave rate. The microstructure is important to heave rate in that the shape of the ice-water interphase is expected to have some control over the suction pressure. As mentioned in Chapter II, the fact that the ice within the soil pores tends to fill the crevices of the pores at colder temperatures (and higher pressures), thus reducing the apparent permeability, could contribute to this time dependent behavior. The deformation of ice in the soil pores has not be fully examined. In this regard, the proliferation of ice into the ice free pores may have occurred in time reducing the permeability of the soil. The exact nature of these processes and their effect on heave rate is still not clear.

In conclusion, the fact that heave rate decreases with time is generally accepted. Although the reasons for this behavior are not well understood it is suggested that the decrease in temperature gradient within the frozen zone and the increase in overburden pressure with the depth of frost penetration contribute most to the reduction of heave rate. The boundary and internal conditions of laboratory tests should be similar to that expected in the field in order to use the results to explain field behavior. Further controlled laboratory and especially field testing is required for a better assessment of the factors influencing this process.

# CHAPTER VI

# Conclusions and Recommendations

## 6.1 Conclusions

The purpose of this research was to investigate the phenomenon of moisture migration in frozen soil as it relates to frost heave. In light of the laboratory test results and discussions presented in the previous chapters, the following conclusions can be drawn.

1.) Moisture will migrate through frozen soil under the influence of a temperature gradient. The rate of migration is apparently controlled by the actual temperature distribution (the temperature governs the potential of the unfrozen water) and frozen permeability within the frozen fringe behind the 0°C isotherm.

2.) The permeability of frozen soil is a function of the amount of unfrozen water and of the amount of ice within the soil pores. Moisture will flow with relative ease through the unfrozen water layer of a frozen soil prior to the development of an ice structure. As ice accumulates within the frozen zone

the apparent permeability decreases due to a net increase in tortuosity, i.e., the water must travel a greater distance around the ice structure in order to move up through the soil.

3.) One effect of overburden pressure on a freezing soil-water system is to create a larger zone of ice accumulation. This observed behavior supports the prediction of Penner and Walton (1978).

4.) It appears that the rate of frost heave in a freezing soil is controlled only by the conditions within a relatively thin zone of frozen soil near the frozen-unfrozen interphase termed the "frozen fringe". These conditions include such factors as temperature gradient, overburden pressure and void ratio. The depth to the water table affects the heave rate indirectly. The interrelationship between these parameters is not completely understood at this time. The important concept is that the suction within the frozen soil behind the warmest ice lense contributes little if anything to the rate of heave.

5.) The process of regelation (the movement of pore ice through a frozen soil) occurs in a frozen soil as a result of an applied temperature gradient. However, it appears that this process is very slow and

contributes little to the rate of heave within an engineering time frame when compared with the effect of suction within the frozen fringe.

## 6.2 Recommendations

From the conclusions it can be seen that the phenomenon of frost heave is primarily an interphase problem. It is suggested that future research be directed towards the examination of the frozen fringe of soil between the warmest ice lense and the 0°C isotherm. In particular, it is of interest to investigate the effects of temperature, particle configuration (void ratio) and pressure on the magnitude of suction pressure within this zone. Precise temperature measurements and a means of accurately identifying the actual frozen-unfrozen boundary would be required to obtain the necessary information for relating experimental results to thermodynamics. It would be useful to obtain an understanding of exactly where in this fringe the suction is developed and which factors most influence its magnitude.

One of the more difficult problems expected to be encountered with this approach would be the non-uniformity of pore sizes. This problem could be circumvented by the development of an apparatus consisting of many uniform,

precise capillary tubes of a known diameter. Frost heave experiments would be conducted within these tubes and, provided good temperature measurements can be obtained, a model based on thermodynamic relationships could be developed.

From this model certain key parameters would result. The parameters would include the effect of such influential factors as temperature, soil type (surface area), pressure, void ratio (or permeability) and time. The model could be used to predict heave characteristics in a soil by relating the parameters obtained from the control test to those obtained from frost heave tests on the soil.

It is stressed that any model derived from laboratory testing should not be considered for engineering design until the model has been adequately tested with quality data from field frost heave tests. It is important to study the behaviour of a freezing soil system under field conditions in order to verify that the processes assumed in the development of the model actually occur.

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# APPENDIX A

## Conversion Table and List of Equipment

# FIGURE A.1

LIST OF RECOMMENDED UNITS, UNIT ABBREVIATIONS, QUANTITY SYMBOLS AND CONVERSION FACTORS FOR USE IN GEOTECHNICAL ENGINEERING

SI BASE UNITS, DERIVED UNITS AND MULTIPLES

| QUANTITY AND (SYMBOL) | UNITS AND MULTIPLES | UNIT ABBREVIATION | CONVERSION FACTORS FOR EXISTING UNITS | REMARKS |
|---|---|---|---|---|
| Length (various) | kilometer<br>meter<br>centimeter<br>millimeter<br>micrometer | km<br>m<br>cm<br>mm<br>μm | 1 mile = 1.609 km<br>1 foot = 0.3048 m<br>1 inch = 2.540 cm<br>1 inch = 25.40 mm | 1 micrometer = 1 micron |
| Area (A) | square kilometer<br>square meter<br>square centimeter<br>square millimeter | $km^2$<br>$m^2$<br>$cm^2$<br>$mm^2$ | 1 $mile^2$ = 2.590 $km^2$<br>1 $foot^2$ = 0.09290 $m^2$<br>1 $inch^2$ = 6.452 $cm^2$<br>1 $inch^2$ = 645.2 $mm^2$ | 1 acre = 0.4047 hectares<br>1 hectare = 10,000 $m^2$ |
| Volume (V) | cubic meter<br>cubic centimeter<br>cubic millimeter | $m^3$<br>$cm^3$<br>$mm^3$ | 1 $foot^3$ = 0.02832 $m^3$<br>1 $inch^3$ = 16.39 $cm^3$<br>1 Imp. gallon = 4546 $cm^3$ | To be used for solids and liquids |
| Mass (m) | megagram (or tonne)<br>kilogram<br>gram | Mg<br>kg<br>g | 1 ton = 1.016 Mg<br>1 lb = 0.4536 kg<br>1 lb = 453.6 g | Megagram is the SI term |
| Density (mass density) (ρ) | megagram (or tonne) per cubic meter | $Mg/m^3$<br>$(t/m^3)$ | 100 $lb/ft^3$ = 1.602 $Mg/m^3$<br>(62.43 $lb/ft^3$ pure water = 1 $Mg/m^3$ = spec. grav. 1.0 approx.) | Density is mass per unit volume |
| Force (various) | meganewton<br>kilonewton<br>newton | MN<br>kN<br>N | 1 tonf = 9.964 KN<br>1 lbf = 4.448 N | |
| Weight (force) density ($\gamma = \rho g$) where g is the local gravitational acceleration | kilonewton per cubic meter | $kN/m^3$ | If g = 9.807 $m/s^2$ then a substance with a mass density of 1.0 $Mg/m^3$ has a weight density of 9.807 $KN/m^3$ | No explicit SI unit or term for weight or weight density |
| Pressure (p, u) | megapascal<br>(meganewton per square meter) | MPa<br>$(MN/m^2)$ | 1 $tonf/m^2$ = 15.44 MPa | To be used for shear strength, compressive strength, bearing capacity, elastic moduli and laboratory pressures for rock |
| Stress (σ,t)<br>and<br>Elastic modulii (E,G,K) | kilopascal<br>(kilonewton per square meter) | kPa<br>$(kN/m^2)$ | 1 $lbf/in^2$ = 6.895 KPa<br>1 $lbf/ft^2$ = 0.04788 kPa<br>1 $tonf/ft^2$ = 107.3 kPa<br>1 $kgm/cm^2$ = 97.86 kPa | Ditto for soils |
| Coefficient of volume compressibility ($m_v$) or swelling ($m_s$) | square meter per meganewton or square meter per kilonewton | $m^2/MN$ $(m^2MPa^{-1})$<br>$m^2/KN$ $(m^2kPa^{-1})$ | 1 $ft^2/tonf$ = 9.324 $m^2/MN$<br>= 9.324 x $10^{-3}$ $m^2/kN$ | |
| Coefficient of water permeability $k_w$ | meters per second | m/s | 1 cm/s = 0.010 m/s<br>1 ft/d = 3.53 x $10^{-6}$ m/s | This is a velocity depending on temperature and defined by Darcy's law<br>$V = k_w \frac{\partial h}{\partial s}$<br>where $\partial h/\partial s$ = hydraulic gradient |
| Coefficient of consolidation ($c_v$) or swelling ($c_s$)<br>$c_v = k/\gamma_w m_v$ | square meter per year | $m^2$/year | 1 $ft^2/yr$ = 0.09290 $m^2/yr$<br>1 $ft^2/day$ = 33.95 $m^2/yr$<br>1 $cm^2/s$ = 3.154 x $10^3$ $m^2/yr$ | |
| Absolute permeability | square micrometer | $mm^2$ | 1 Darcy = 0.9869 $mm^2$ | This is an area which quantifies the seepage properties of a soil independent of the fluid concerned<br>$v = \frac{k\rho g}{\eta} \frac{\partial h}{\partial s}$<br>where ρ = fluid density<br>g = gravitational acceleration<br>η = dynamic viscosity |
| Dynamic viscosity (η) | millipascal second<br>(centipoise) | mPas<br>(cP) | 1 cP = 1 mPas | Dyanmic viscosity is defined by Stoke's Law |
| Kinematic viscosity (ν) | square millimeter per second<br>(centistoke) | $mm^2/s$<br>(cSt) | 1 cSt = 1 $mm^2/s$ | $\nu = \eta/\rho$ |
| Celcius temperature (T,θ) | degree Celcius | °C | T °F = 5 (T-32)/9 °C | 0°C = 273.15 °K |
| Heat | Joule | J | 1 cal = 4.19 J<br>1 BTU = 1056 J | |
| Heat flux | watt per square meter | $W/m^2$ | 1 $cal/cm^2/s$ = 4.19 x $10^4$ $W/m^2$ | |
| Coefficient of thermal conductivity ($k_u$,$k_f$) | watt per meter degree K (or C) | W/m.K (W/m.C) | 1 cal/cm.s.°C = 4.19 x $10^2$ W/m.K<br>1 BTU/hr.ft.°F = 1.730 W/m.K | |
| Plane angle (various) | degree<br>minute<br>second (angle) | °<br>'<br>'' | | Used for angle of shearing resistance (ϕ) and for slopes |
| Time (t) | year<br>day<br>hour<br>second (time) | year<br>d<br>h<br>s | 1 year = 31.56 x $10^6$ s<br>1d = 86.40 x $10^3$ s<br>1h = 3600s | 'a' is the abbreviation for year<br>The second is the SI unit |

FIGURE A.2

List of Equipment

HOTPACK - constant temperature/circulating bath
Model No. 334

Available from Hotpack (Canada) Ltd. 385 Phillip Street North, Waterloo, Ontario

PACE Wiancko - 0 to 0.5 psi pressure transducer
Model P.I.D.-0.5 psid

Available from Pace-Wancko, Division of Whittaker, N. Hollywood, California.

HEWLETT-PACKARD - 6 Volt excitation L.V.D.T.
Model 7DCDT-500

Available from Hewlett-Packard, Waltham, Mass.

ATKINS - Resistance Thermistors
Model PR-3

Available from York Instruments, 1262 Don Mills Road, Don Mills, Ontario
or from Atkins Technical Inc.,Box 14405 University Station, Stengel Airfield, Gainesville, Florida 32603

FLUKE - data aquisition system
Model No. 2240A

Available from John Fluke Mfg. Co. Ltd., P.O. Box 43210, Mountlake Terrace, Washington 98043

TECHTRAN - cassette recording unit
Model No. 8400/8410

Available from Techtran Industries Inc.,
580 Jefferson Road, Rochester, New York 14623

APPENDIX B

Plots of Test Results

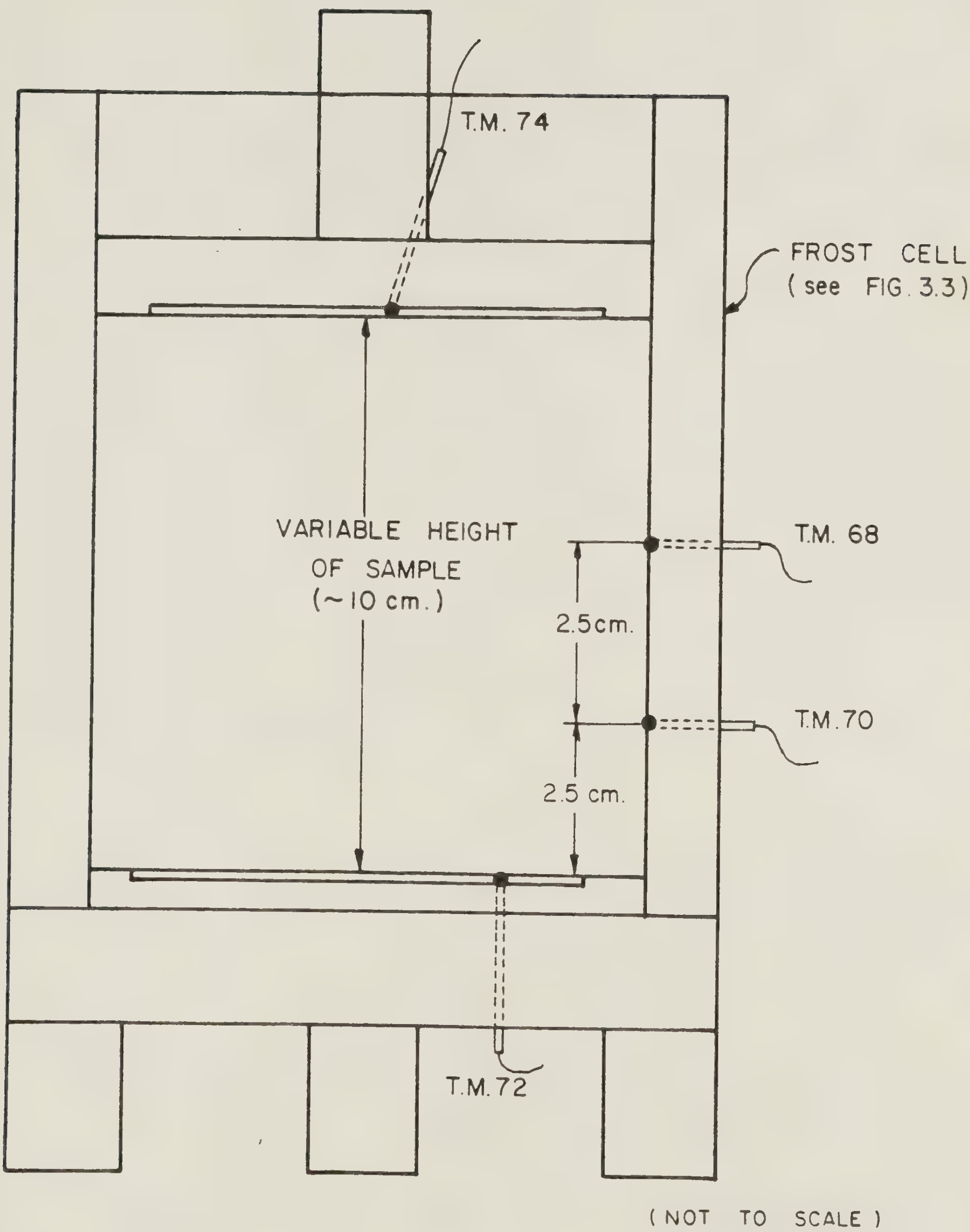
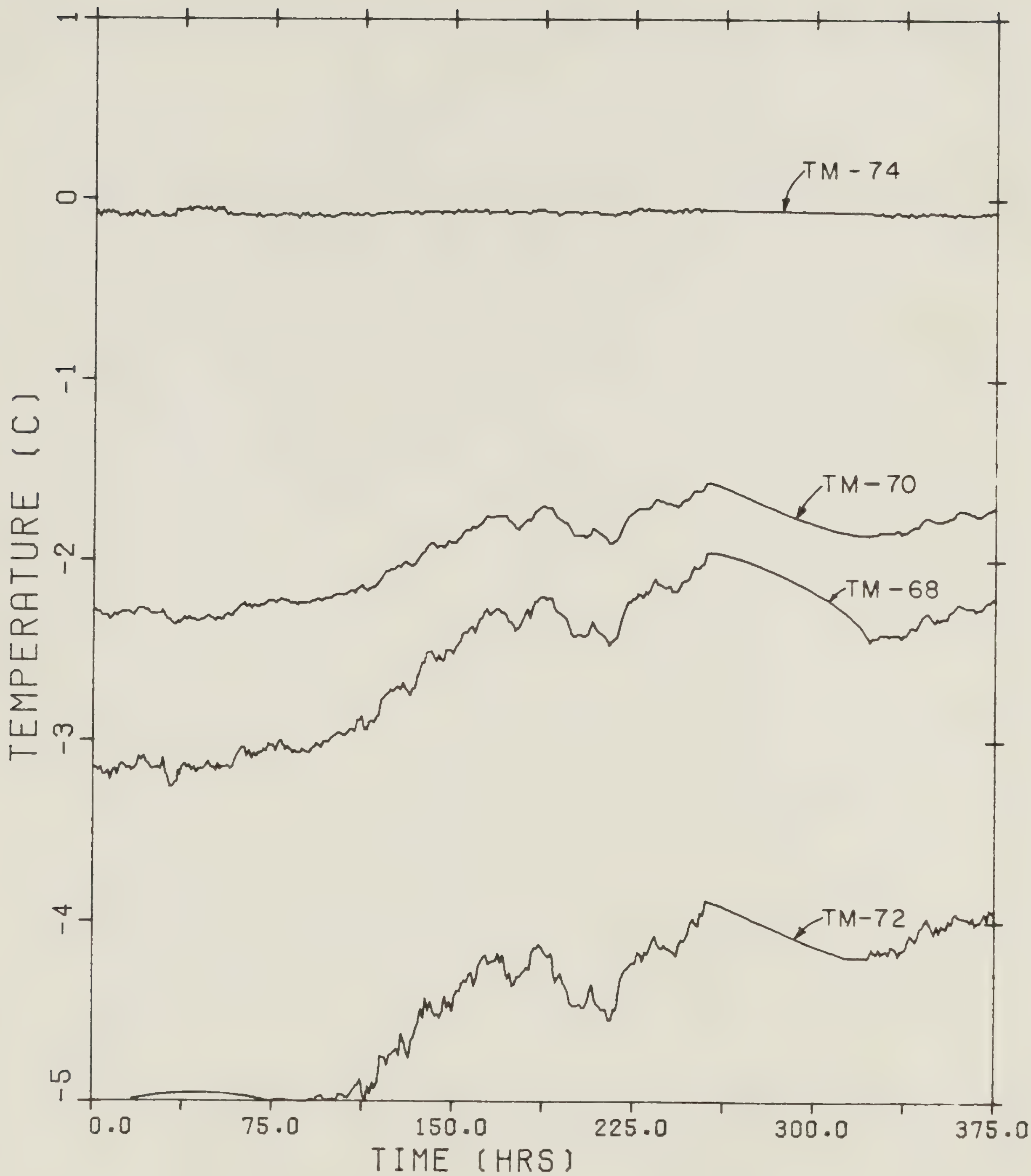
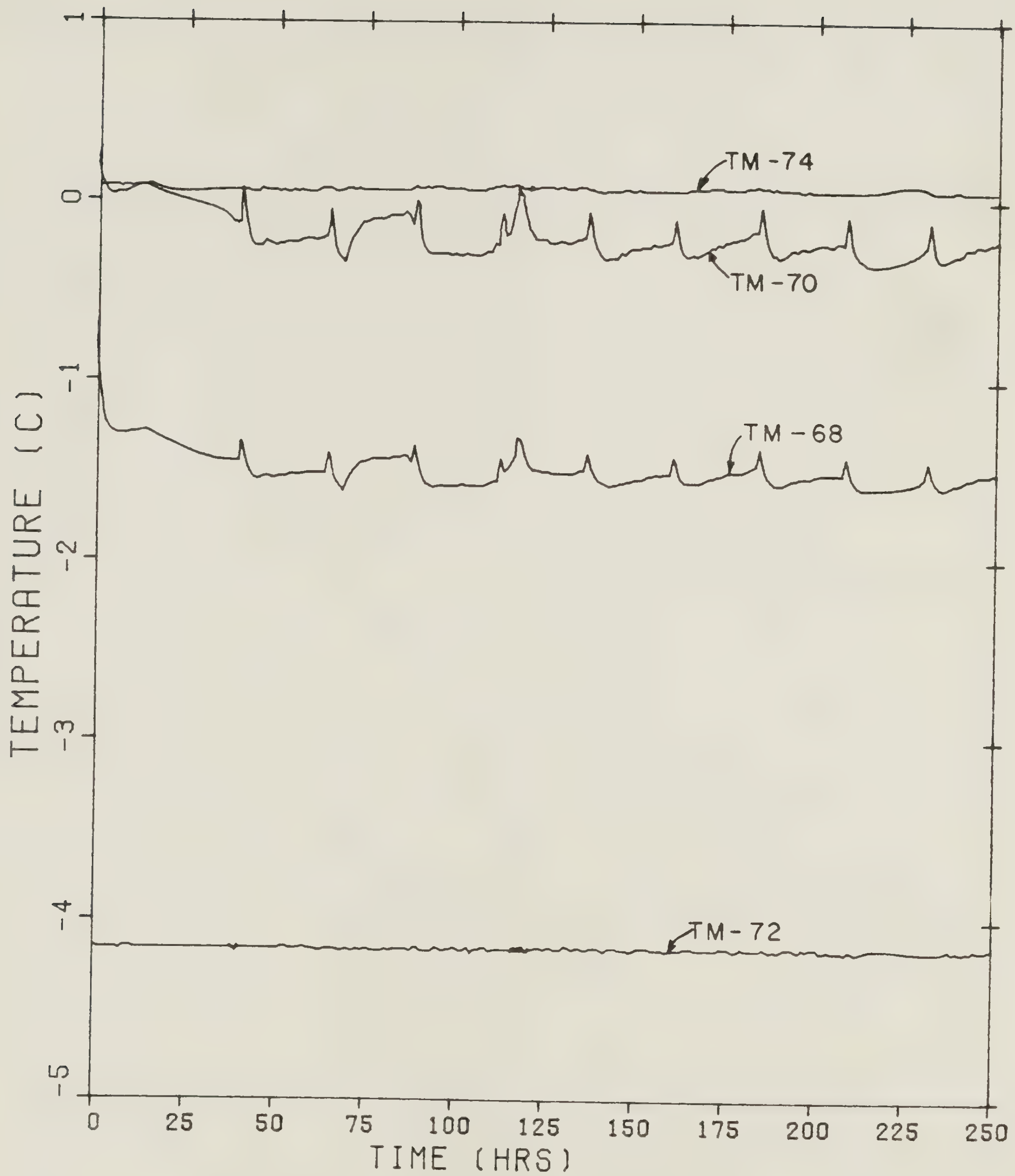



FIGURE B-1 : SCHEMATIC DIAGRAM OF THERMISTOR POSITIONS.



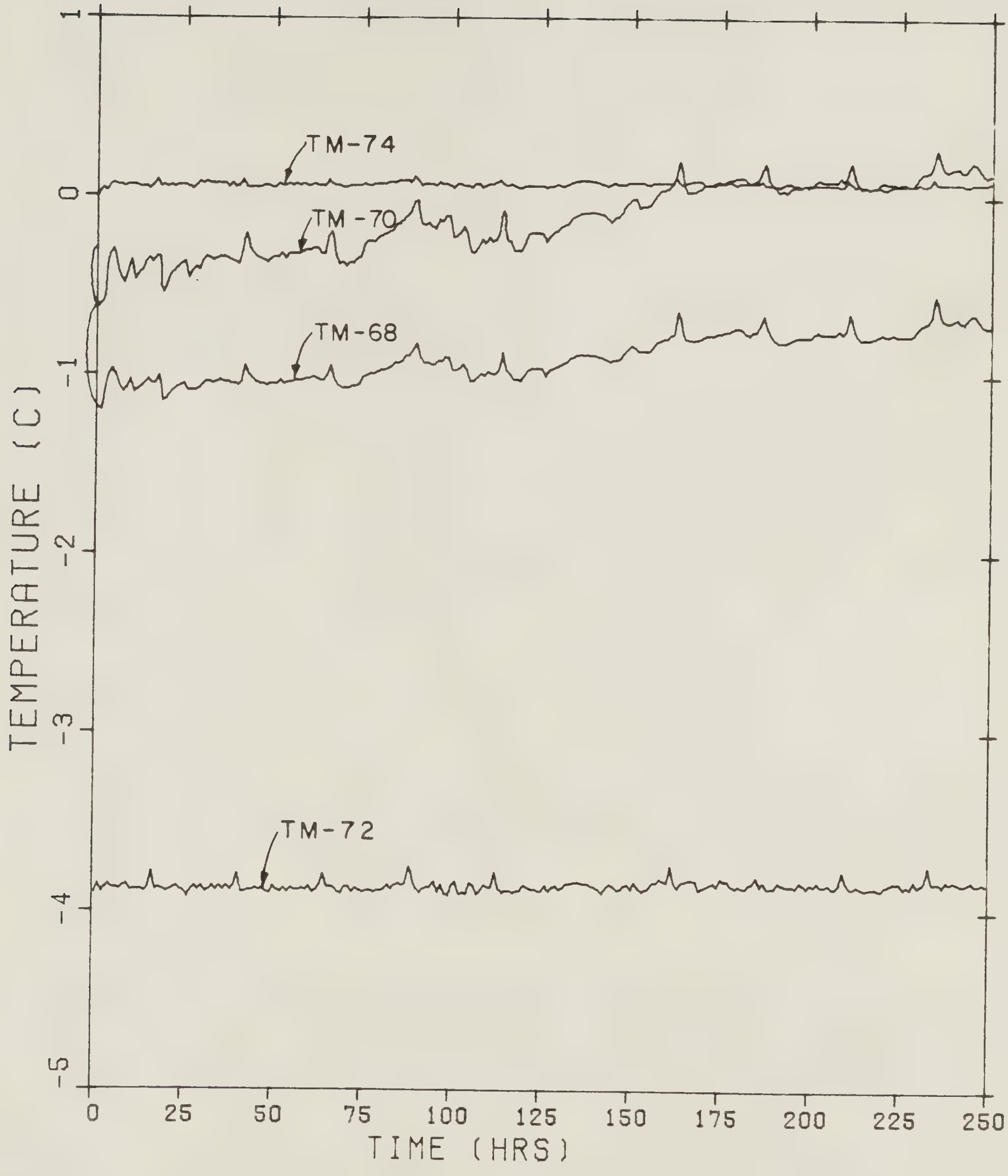

TEST A-2 (P=0.00;CLOSED SYSTEM)

FIGURE B-2



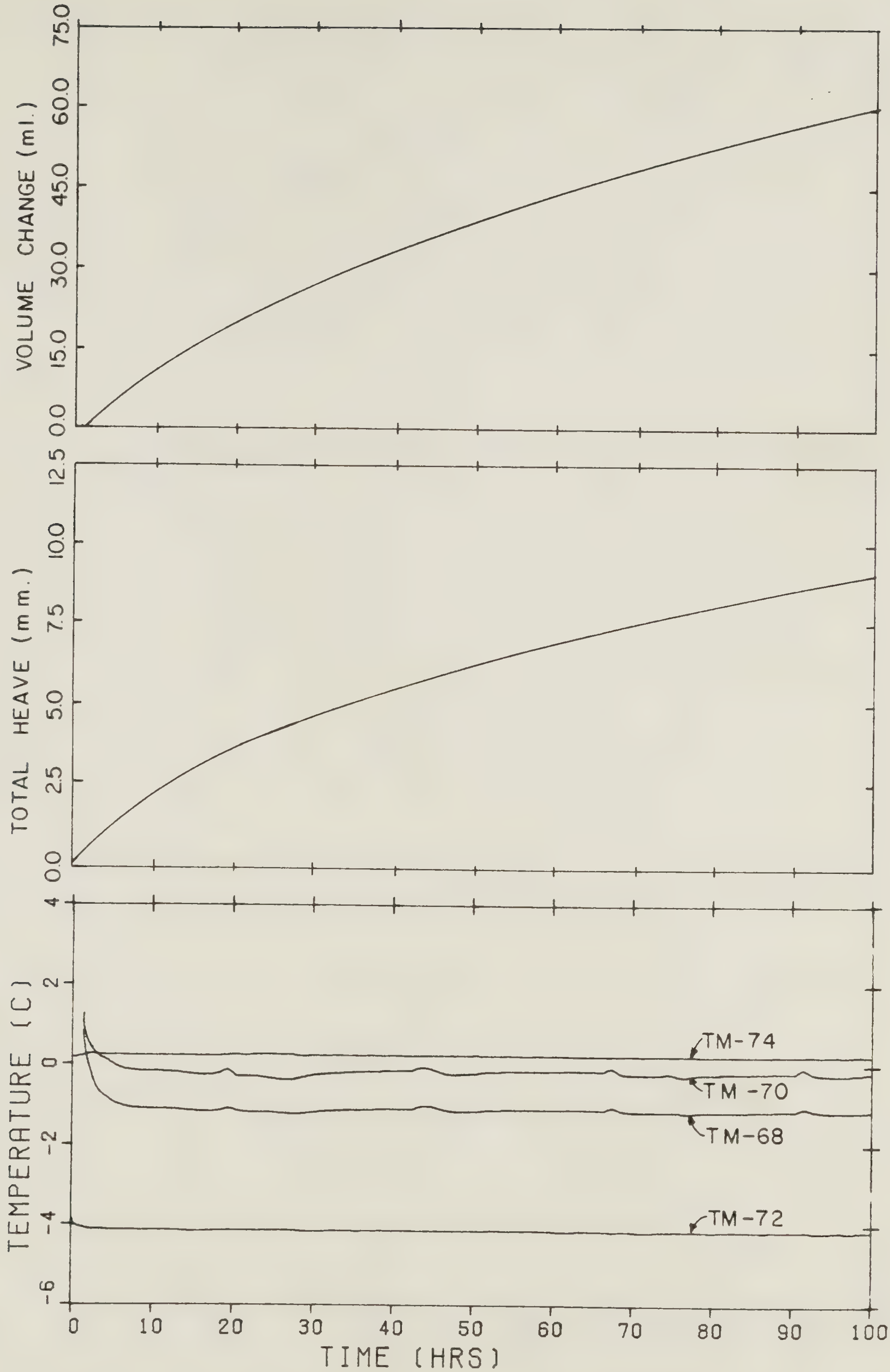

TEST A-3 (P=0.00;CLOSED SYSTEM)

FIGURE B-3



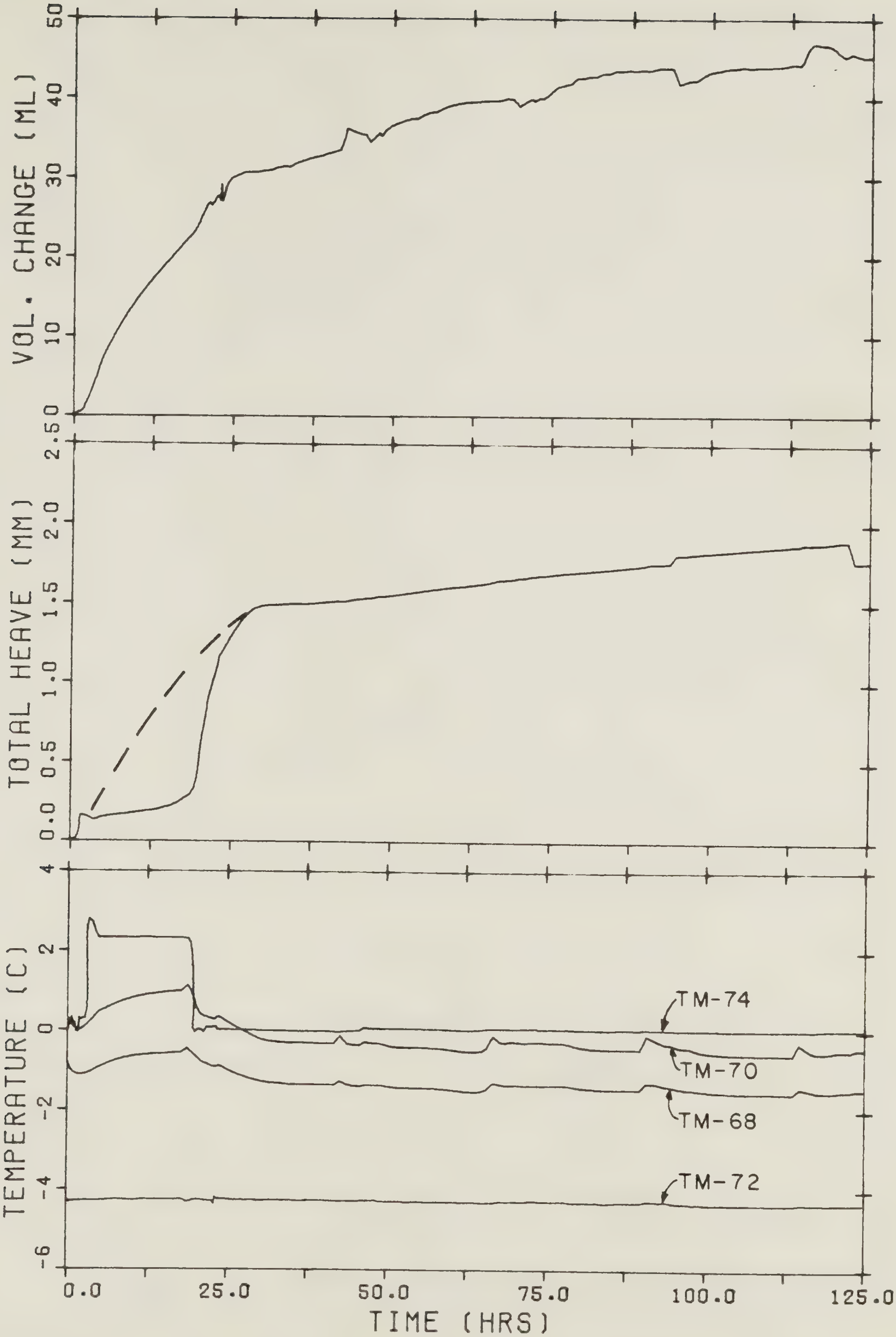

TEST A-4 (P=0.00;CLOSED SYSTEM)

FIGURE B-4



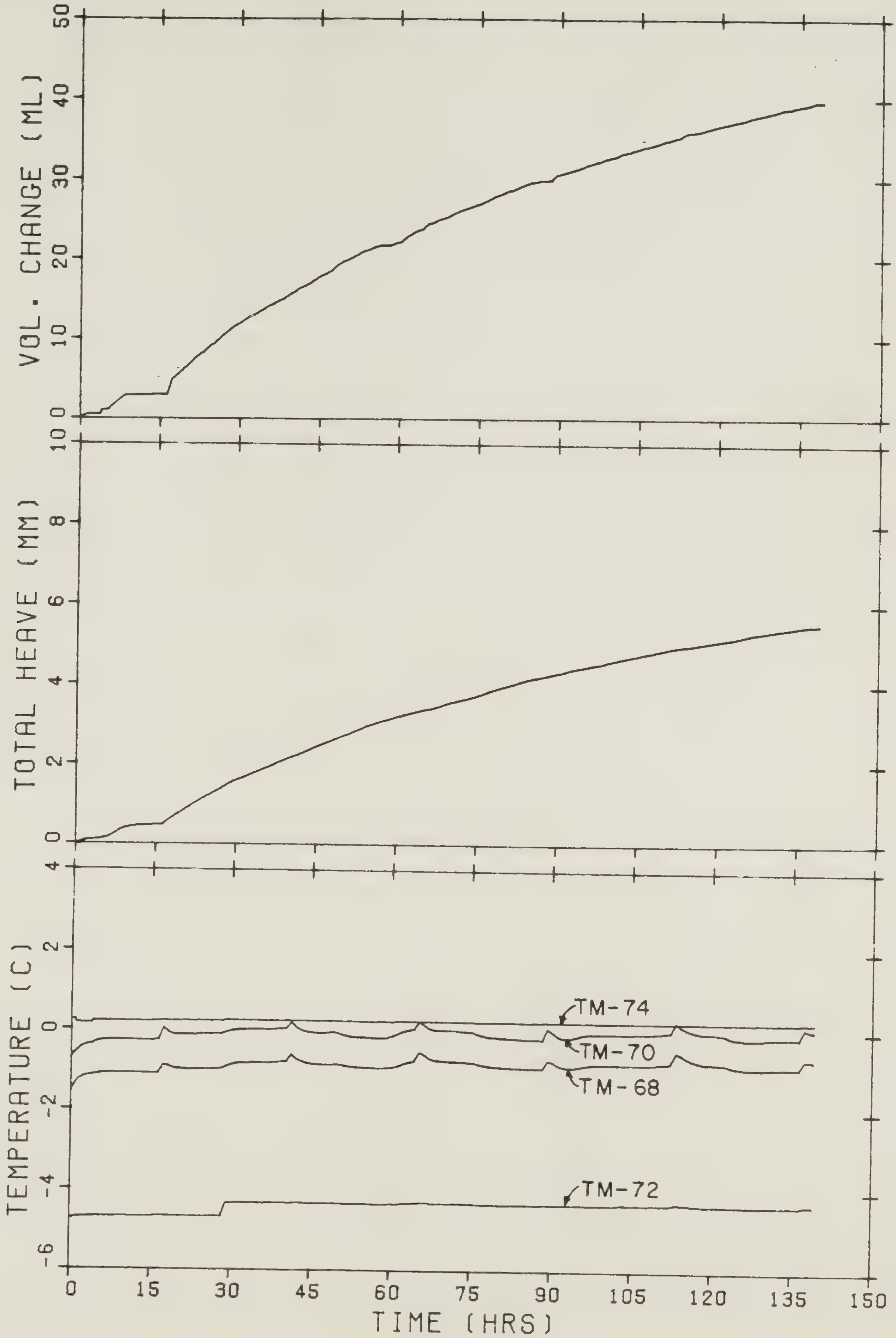

TEST A-5 (P=0.00; OPEN SYSTEM)

FIGURE B-5



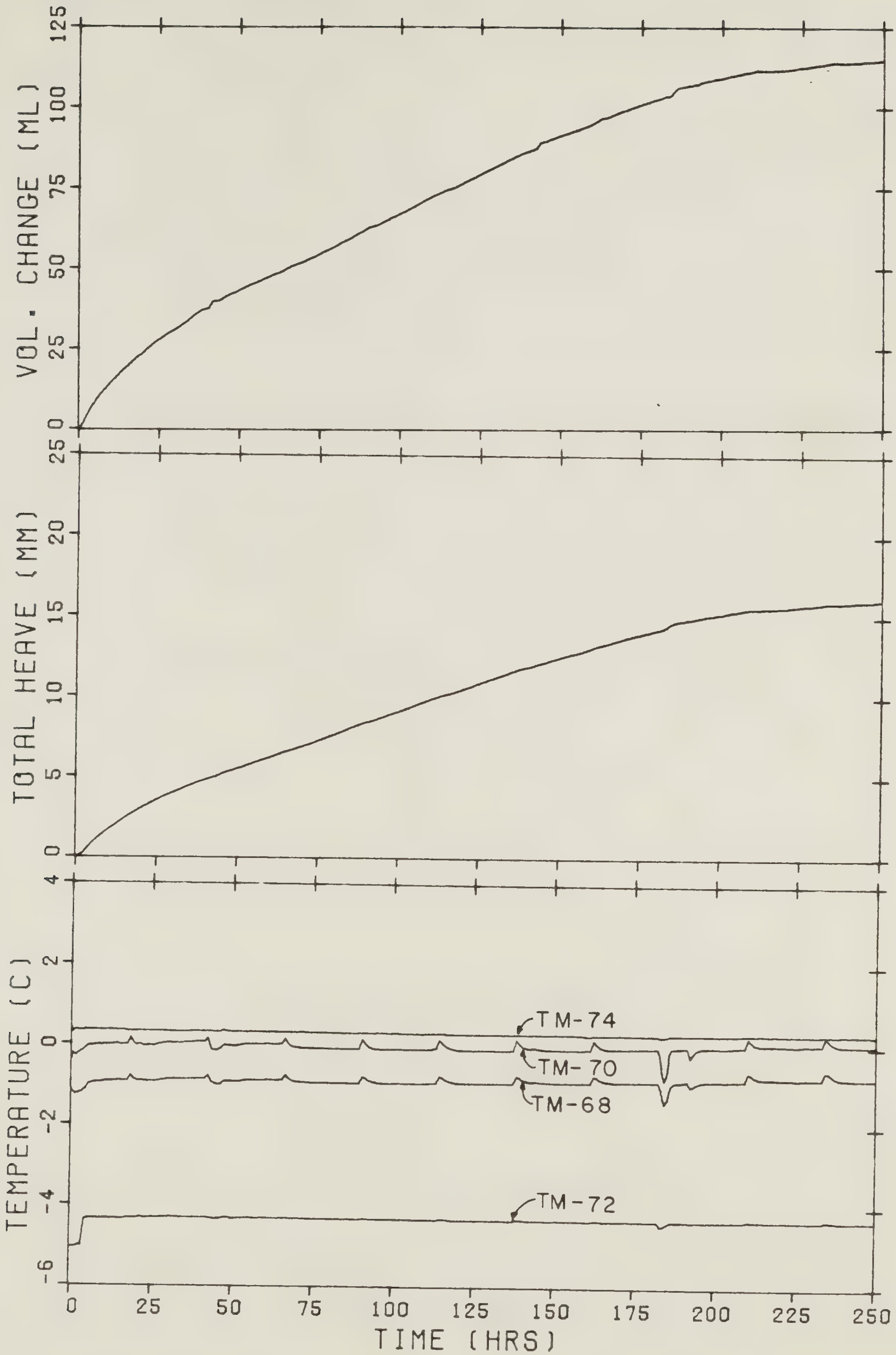

TEST A-6 (P=0.00; OPEN SYSTEM)

FIGURE B-6



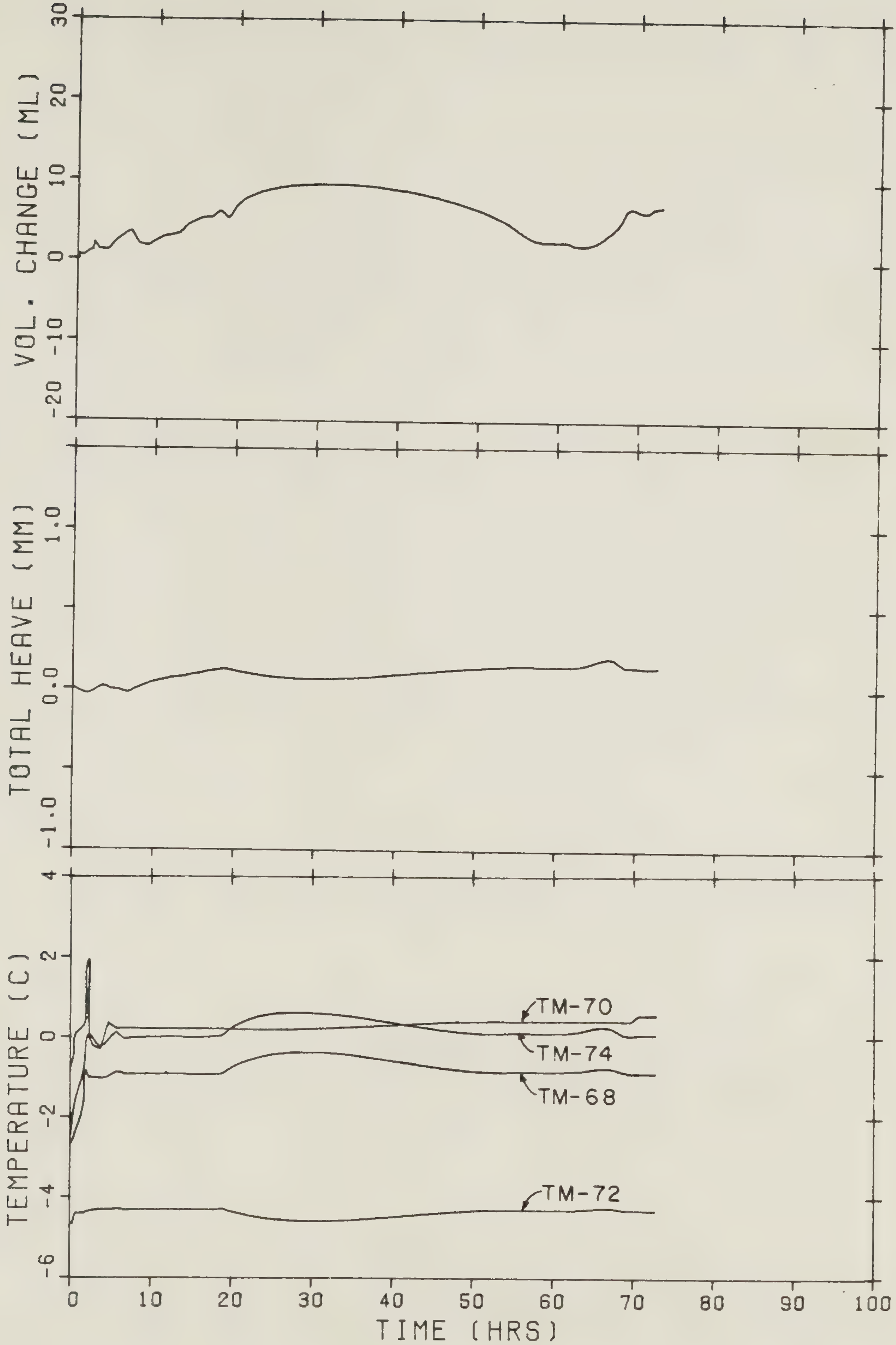

TEST A-7 (P=0.00; OPEN SYSTEM)

FIGURE B-7



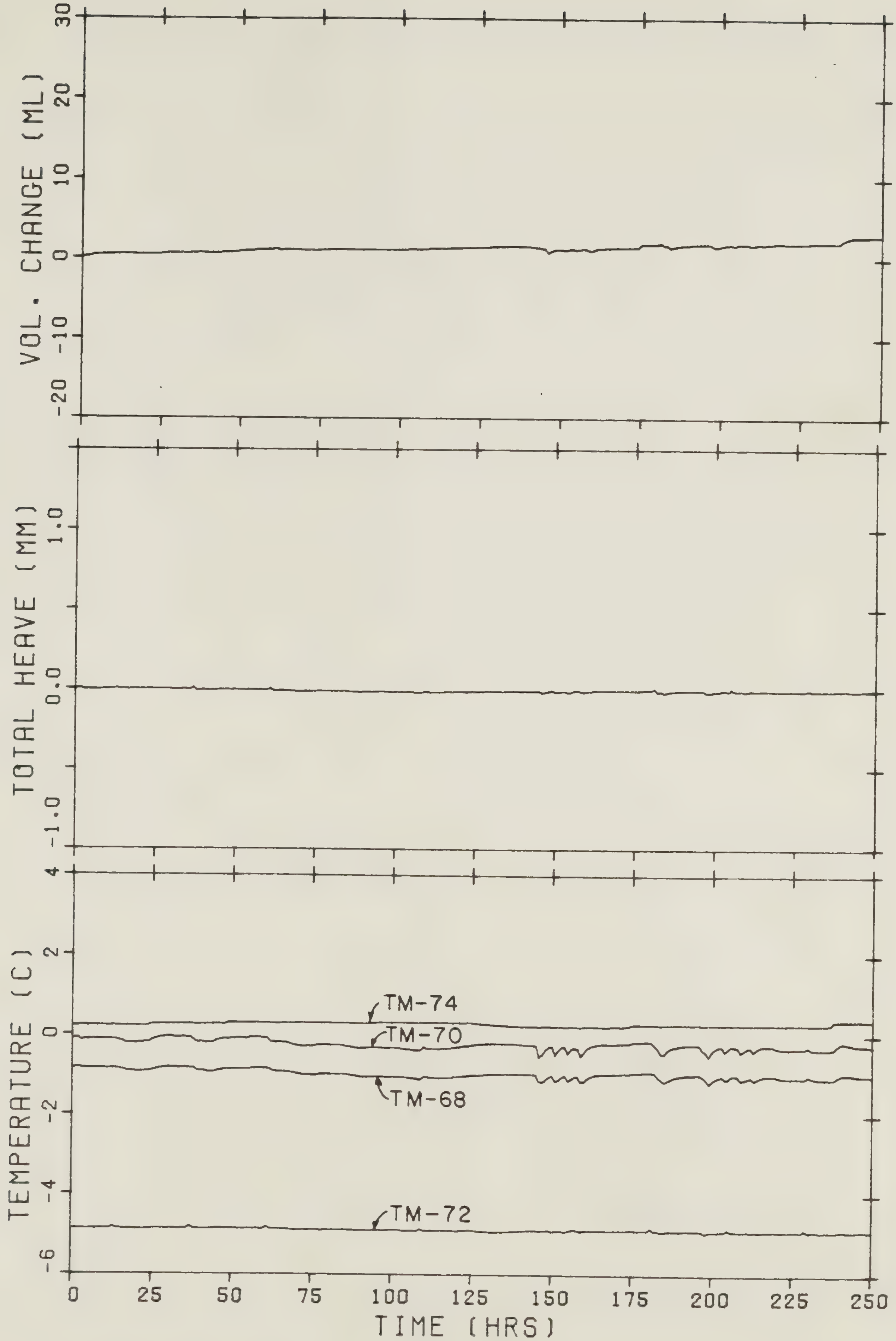

TEST A-8 (P=0.00; OPEN SYSTEM)

**FIGURE B-8**



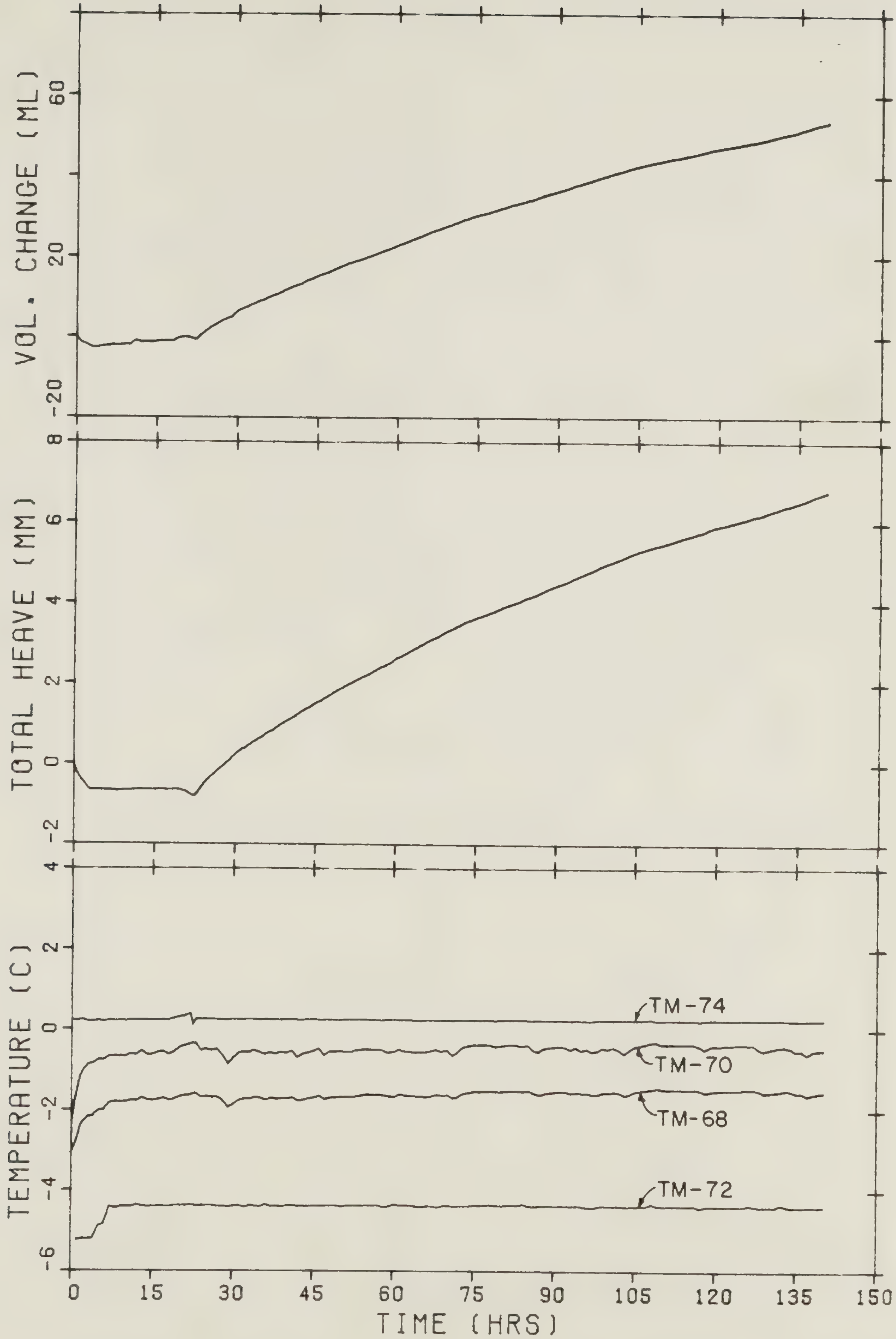

TEST A-9A (P=0.00; OPEN SYSTEM)

FIGURE B-9A



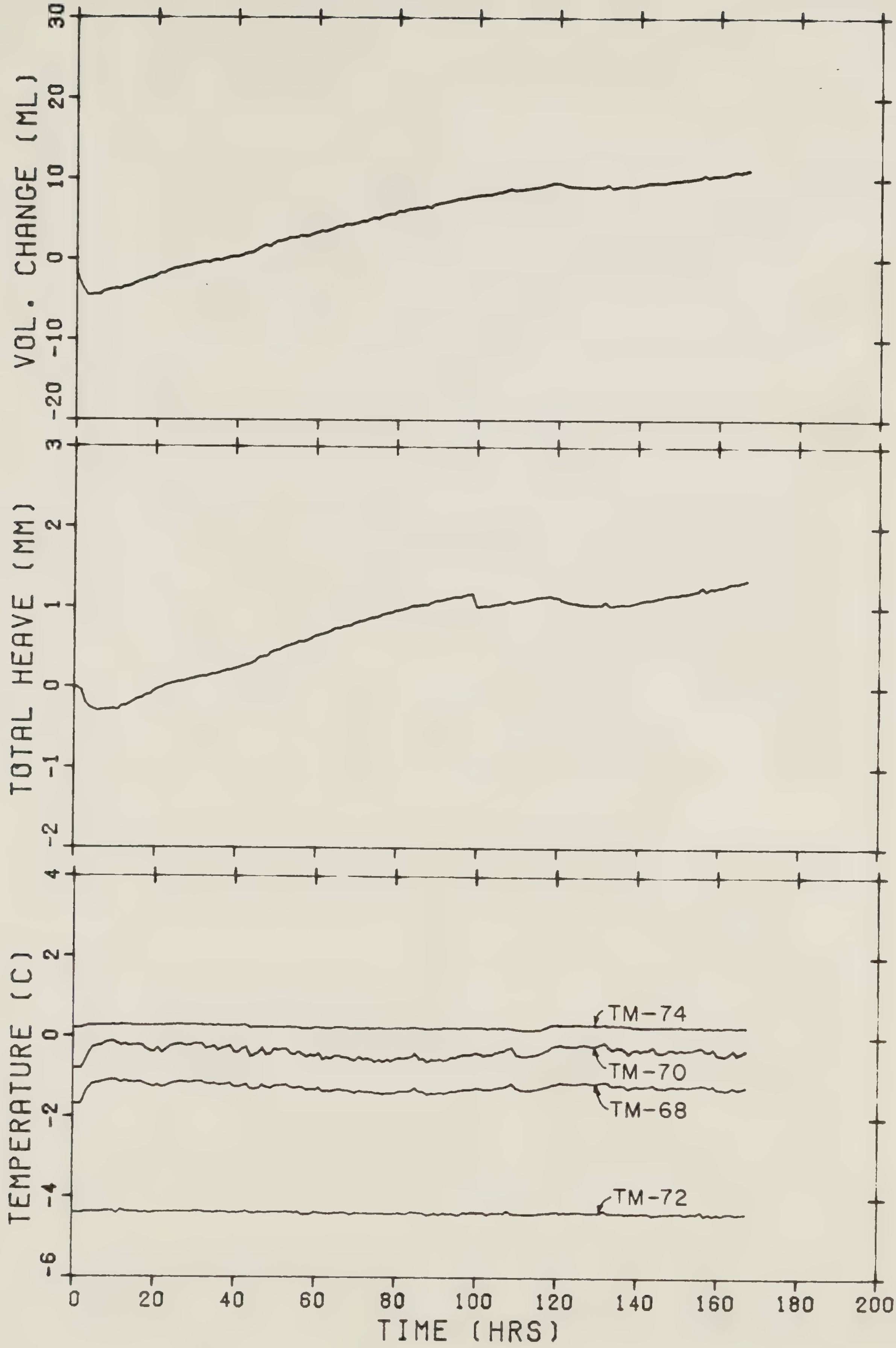

TEST A-9B (P=0.00; OPEN SYSTEM)

FIGURE B-9B



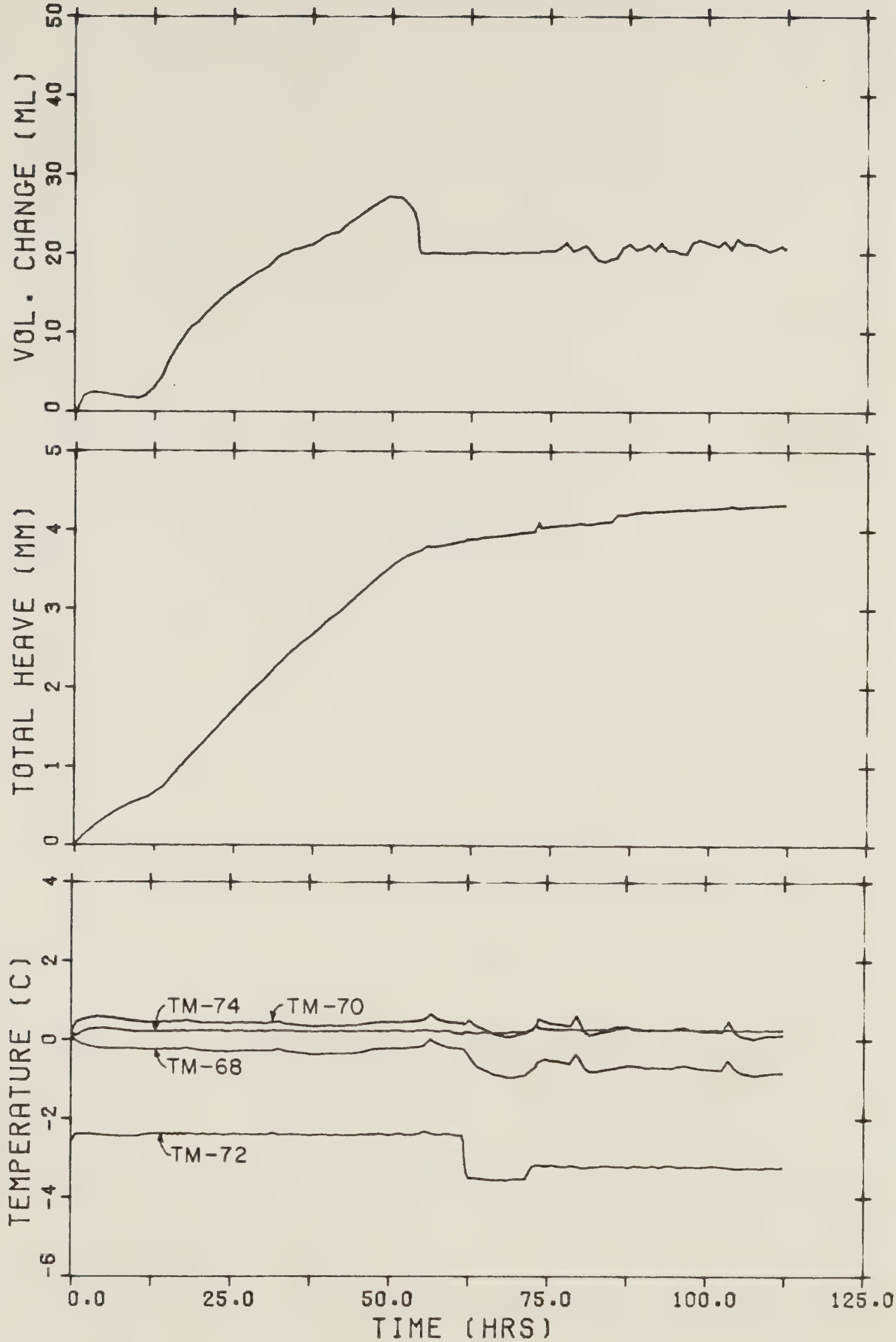

TEST A-10 (P= 50 KPa; OPEN SYSTEM)

FIGURE B-10



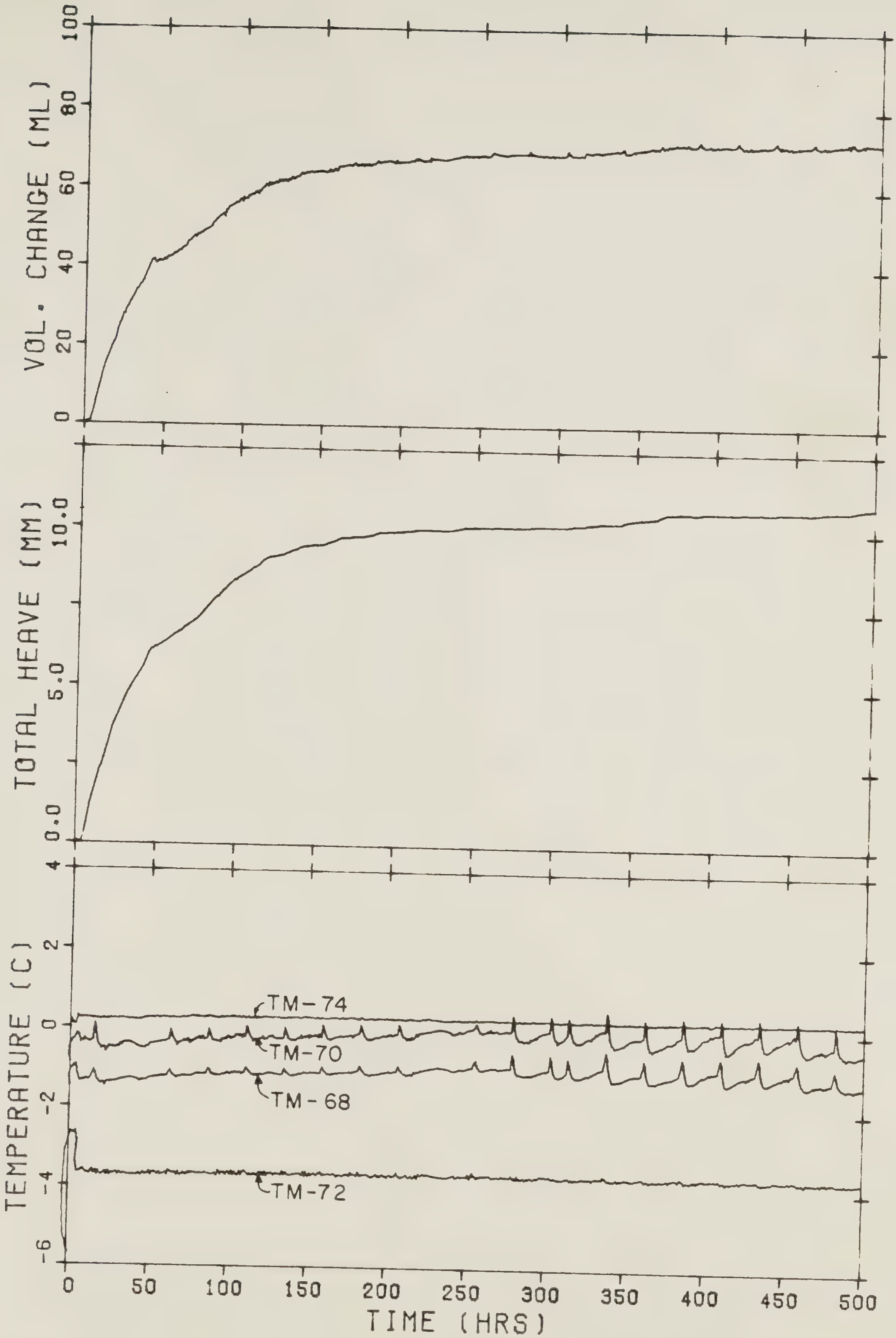

TEST A-11 (P=75KPA; OPEN SYSTEM)

FIGURE B-11




TEST A-12 (P=0.0; OPEN SYSTEM)

FIGURE B-12




TEST A-13 (P=0.0; OPEN SYSTEM)

FIGURE B-13